PHYSICO-GEOGRAPHICAL RESEARCH IN HUNGARY

Geographical Research Institute
Research Centre for Earth Sciences
Hungarian Academy of Sciences
Budapest 2000

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STUDIES IN GEOGRAPHY IN HUNGARY, 32

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PHYSICO-GEOGRAPHICAL RESEARCH IN HUNGARY

The publication of the present volume was supported by the National Scientific Research Fund (OTKA), Project No. T 034730.

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Geographical Research Institute
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Hungarian Academy of Sciences

GEOGRAPHICAL RESEARCH INSTITUTE HAS
BUDAPEST 2000

Preparation for printing carried out at the
Geographical Research Institute
Research Centre for Earth Sciences
Hungarian Academy of Sciences (GRI RCES HAS)

Computer typesetting and layout:
ESZTER GARAI

Technical editor:
LÁSZLÓ BASSA

Technical staff:
ALEXANDRA KOVÁCS, ISTVÁN POÓR

ISBN 963 7395 83 0
HU ISSN 0081-7961

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Published by Geographical Research Institute Research Centre for Earth Sciences
Hungarian Academy of Sciences, H-1112 Budapest, Budaörsi út 43-45.

Printed in Hungary
by EXEON Bt.

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Preface

Nowadays the trend of globalisation is absorbing all spheres of life. The cumulative environmental problems, the expansion of tourism, the information boom all have contributed to an ever-increasing interest of the public and academic audience toward earth sciences. This issues a real challenge to geography which is trying to take it up and respond to it by the development of new methods and by the application of the recent scientific results for the solution of functional problems of society and economy. It does not mean, naturally, that the 'classical' branches of geography wither and the related fundamental research have been neglected because the solutions of practical tasks are increasingly based upon them.

This volume of studies is aimed at the presentation of the state-of-the-art of research activities in physical geographical studies in Hungary and achievements over the last few years. The contributions have been selected with the purpose that the geographical workshops in Hungary be represented and beyond the geomorphologic domain being traditional in the Hungarian physical geography, other and more recent subjects such as landscape ecology and management, GIS applications be presented. The studies were grouped into four chapters according to their thematic content.

In the first chapter some of the results in the field of geochronology were presented. Within this topic the late Neogene has long been a cardinal problem. It was put in the focus of the study by F. SCHWEITZER deserving a special attention. The author has dealt with evidence to the past tropical arid climate in the Carpathian Basin simultaneous with the formation of Mediterranean Upper Miocene *Messinian* strata, thus reformulating the chronology of the Hungarian landscape evolution. The article offers a synthesis of the related knowledge so far accumulated by various branches of geosciences. Quaternary studies being a paramount subject in Hungary also belong to the latter. M. PÉCSI and colleagues report on terrace morphological and loess chronological investigations in the Yenisey valley under the aegis of global correlation. Two contributions deal with environmental changes during the late Pleistocene and Holocene. Of them the first one (by Gy. LOVÁSZ) treats possible blown sand formation and landslide occurrence by Holocene phases. Gy. GÁBRIS and colleagues have modified the former Quaternary stratigraphy in Hungary. The geomorphological analysis of sequences comprising loess, sand and paleosols supported by the luminescence dating method has led to a conclusion (among others) that there was a dry interval (even if a short one) during the Atlantic, about 6000 years ago. These new result contradict the former concept about Holocene, considering the Atlantic the most humid phase of that epoch. TL and IRSL, recently used in loess-paleosol stratigraphy has met with skepticism by many experts as reliable dating methods. At the same time it is a fact that once a new scientific result does not fit in the hitherto accepted concept it should overcome barriers to become recognised.

GIS and different kinds of digital models have recently become widespread also in Hungarian geography. The second chapter is an evidence to this trend. Á. KERTÉSZ and colleagues provide soil erosion assessment by one of the latest computer models. An essay by Sz. SZABÓ and A. KERÉNYI also deals with a typical environmental problem, soil acidification. The mapping of this process within a test area is solved using a GIS software. GIS is a correct method in geomorphologic research as it exemplified by the article of R. KISS and G. MEZŐSI. The authors explore possible reconstructions of past landscape evolution within a model area including changes in the drainage network and catchment, with a special reference to variations in erodibility of the different parts. The method is based on the postulate that for each *Horton-Strahler* drainage system the absent volume i.e. the vertical dissection can be defined (by GIS, models, DTM). Knowing the geological history of the area, the evolution of the watershed can be calculated. The article provides new evidence to temporal changes in river systems through the analysis of data relating to streams.

The third chapter is devoted to karst research as one of the most developed branches in Hungarian geography and probably the most intrinsic subject of the latter. This is well represented by two contributions of L. JAKUCS and I. KEVEI-BÁRÁNY in which landscape ecological approach and the ideas of complexity prevail. In a case study by M. VERESS the problems of geomorphological reconstruction and differentiation of the paleokarsts are tackled and methods of investigation into the cover strata are described. The study of karst can also be used in the reconstruction of the Quaternary. It is well exemplified by the article of L. ZÁMBÓ et al which besides others, points out that the karstic processes events by the nature of their complexity are highly dependent on climate and are especially sensitive to environmental changes. This way the speleothem is an excellent indicator of the changes the paleoenvironment. The authors in compare the result of the investigations performed in the largest allogenic river cave in Hungary, the Baradla, with other paleoclimatic reconstruction methods (quartermalacology, oxygen isotope) as well as the result gained by analogous methods applied in other Northwest European caves and the data obtained are parallel. Interesting conclusions can be drawn for the two above regions in the case if they represent differing characteristic features.

Finally, the fourth chapter contains studies on the interrelationship between landscape evolution and issues of present-day land uses. It begins with J. SZABÓ's case study dealing with the problems of a Hungarian region endangered by landslide hazard and coming to an intriguing conclusion with regard to its modern utilisation. S. MAROSI treated a flood plain area of the Danube and formulated recommendations as to the prospects of its man induced development. D. LÓCZY analysed the overall impact of land privatisation upon Hungarian landscapes and also dealt with the reclamation of areas abandoned by mining. G. HORVÁTH and colleagues wrote an essay, which is a classical landscape description with a substantial attention paid to landforms and processes as a consequence of human intervention.

Recommending this short volume of studies to the reader may we hope that he will be provided an insight into the recent physico-geographical research in Hungary, at the same time getting acquainted with achievements as a result of technical development and the dissemination of new scientific theories and methods on the one hand and expectations of the society on the other hand. Both professionals in geography and a wider audience interested in geosciences can make use of this publication.

The Editors

Á. KERTÉSZ and F. SCHWEITZER (eds): Physico-geographical Research in Hungary
Studies in Geography in Hungary 32, Geographical Research Inst. HAS, Budapest 2000, pp. 9–27.

GEOMORPHIC EVOLUTION IN THE CARPATHIAN BASIN DURING THE LATE CENOZOIC AND THE PLIOCENE EPOCH

FERENC SCHWEITZER[1]

Introduction

The present paper is an attempt to summarise the achievements reached by the author during the past 10–12 years of investigation into the specific features of landform evolution in the Pannonian Basin during the Late Cenozoic and early Quaternary. The major geomorphic processes having taken place at the end of the Upper Miocene and during the Pliocene are also evaluated and landforms described and dated below.

In order to obtain more exact and complex results a special emphasis was put on the application of *biochronological-biostratigraphical methods* based on the identification of mammal fauna on the one hand and of *paleomagnetic datings* (*absolute chronology)* on the other hand.

The formation of the present-day surface of Hungary was preceded by a series of events of several million years' duration comprising the formation of rock sequences studied by the geological science. In the course of Earth history marine, terrestrial and volcanic formations came into existence and land surface varied considerably during the different geological epochs. These paleosurfaces and their geomorphic features are studied by diverse research methods of geosciences through *paleogeographic reconstructions* providing more or less reliable results depending on the amount and quality of geological data.

This retrospective analysis starting with the present is losing its reliability with time because the superimposing geological events eliminate an increasing part of evidence for reconstruction. So, stemming from the character of geological events neither the geomorphological, nor the geological system of data can be complete. This is why geomorphological-geological research always necessitates *hypotheses*. These hypotheses, i.e. the respective theoretical bases, however, should always be well-founded.

Hungary is situated in the Carpathian Basin bounded by the Alps, Carpathians and Dinarids and thus its mountains, basement and sediments are akin to geological formations of the modern mountain structures.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 019326.

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Several recent studies on tectonism and paleogeography conceived in the theory of plate tectonics (GÉCZY, B. 1973, STEGENA, L. 1973, HORVÁTH, F. 1973, MÁRTON, P. and M. SZALAY E. 1978, 1981, BALLA, Z. 1980, 1984, KOVÁCS, S. 1980, 1982, MAJOROS, Gy. 1980, KÁZMÉR, M. 1984, BÁLDI, T. 1982, FÜLÖP, J. *et al.* 1988) came to a joint conclusion that *the Carpathian Basin (and Hungary) as a uniform area have existed approximately since the Middle Miocene.* By this time the area evolved from two lithosphere units: the north-western part of African origin and the south-eastern segment of European origin.

When evaluating the Pliocene epoch a main emphasis was put on general paleogeographic relationships since domestic data are often missing.

It was the absence of data from correlative sediments covering the domestic geomorphological levels that stimulated to approximate their age using analogues from remote locations but well studied by other experts and in several cases by the author himself, One of the greatest tasks of the Hungarian geomorphological research is *to explain and clarify geomorphic evolution during the Pliocene epoch and to propose a paleogeographic reconstruction.*

The duration of the Pliocene, as suggested by the Italian and French experts, is 2.5–3 m yr; in this period 200 to 1300 m thick sediment sequence was deposited over the intra-basin areas while the basin margins and inland areas were covered by sediments in 10 to 250 m thickness (e.g. Gödöllő sand). The epoch restricted in the above sense resulted in an Eastern European terrestrial sequence figuring in the literature as the Levantan stage. Its further subdivision (Piacenzan, Astian), however, was carried out using data from the Mediterranean. Pliocene sediments from the Carpathian Basin do not completely correlate with the originally described Levantan sediments and levels, i.e. there are contradictory concepts about the boundaries.

That is why the denomination 'Levantan' has recently been abandoned by many and the time interval in concern came to be called Upper Pliocene (when the Miocene/Pliocene boundary was placed between the Sarmatian and Pannonian stages) and lately Pliocene stage (with the Miocene/Pliocene boundary drawn at 5.5 m yr).

The Miocene/Pliocene boundary

By the end of the Miocene the Paratethys had been separated into partial basins. One of the basins was the Pannonian brackish inland sea, sediments of which hardly extend beyond the borders of the present-day Hungary.

In the rapidly submerging catchment of the Pannonian Lake a sequence of brackish, lacustrine molasse was deposited in a thickness of 4000 m or more, containing sand, fine grained sand, clay and clayey marl. The lake became entirely filled up by the beginning of the Quaternary.

The Upper Pliocene in Hungary is synonymous with Post-Sarmatian. This stage is placed between the Sarmatian and the Dacian, in the terrestrial biological scale between the Astaracium and Ruscinium, in the mammal system by Mein between zones MN 8 and 12 (STEININGER F.F. *et al.* 1985). Its equivalent in the Hungarian stratigraphical system is the Kunság stage of Pannonian s. str. (DANK, V. and JÁMBOR, Á. 1987). Formations of the Kunság stage contain basal faunas of the Carpathian Basin (type localities Rudabánya, Diósd, Sopron, Csákvár, Győrszentmárton, Sümeg, Szabadság-hegy, Polgárdi, Hatvan, Baltavár etc.) and dispersed findings, biostratigraphical analyses of which were performed chiefly by M. KRETZOI (KRETZOI, M. 1942, 1952, 1961, 1969, 1982, 1987, KORDOS, L. 1987).

Previously the Miocene/Pliocene boundary was drawn in the Carpathian Basin at ca 12 m yr B. P., at the end of the Sarmatian. Lacustrine deposits of the Lower and Upper Pannonian and overlying them freshwater-fluvial, so called Levantan, layers were labelled as Pliocene series.

Recently Lower and Upper Pannonian layers have been claimed by many as of Upper Miocene and the Miocene/Pliocene boundary drawn on top of the Messinian stage (Italy) at 5.3 m yr, whereas Upper Pannonian boundary coincides with that of the Pleistocene (2.5 m yr B.P.). The corresponding sediments are represented in Hungary by the Bérbaltavár sand (KRETZOI, M. 1982).

In the Hungarian stratigraphical practice the *lower and upper boundaries of Pliocene varies therefore it seems to be necessary to write about in what a sense they will be used below.*

The *lower boundary* is drawn, in accordance with the latest international recommendations, between the Messinian and Zanclean at 5.3 m yr, corresponding to the limit between zones MN 13 and MN 14 in the mammal scale by Mein . This was the time of a considerable drop of the level of the Pannonian brackish inland sea, considered a remnant of the Paratethys *(Fig. 1)*, to be correlated with the lowering level of the Mediterranean Sea in the Messinian (5.3–6.8 m yr) with the formation of evaporites in that basin ("Messinian Salinity Crisis"). The reasons for this intense evaporite formation are not yet cleared. Some refer to a cyclic separation of the Mediterranean into parts as a result of plate tectonic movements leading to its drying out. Others hold that there had been a complex relationship between the Mediterranean and the Atlantic interrupted by the following events: a global climate change occurring during the 6^{th} paleomagnetic epoch (6.4 m yr B.P.) under the influence of which the ratio of $\delta^{13}C$ had changed; peak of the Antarctic glaciation (dated 7.4 Ma by the K/Ar method) and a maximum glaciation of Dronning Maud Land (5.5–4.5 Ma K/Ar). At that time glaciation of the Antartic 1.5–2-fold exceeded the present-day one and there was a strong asymmetry between the northern and southern hemispheres (BERGGREN, V. *et al.* 1985, HARLAND *et al.* 1982). The uplift of the Gibraltar swell and the advent of the "Messinian Salinity Crisis" occurred at that time.

The termination of the "Messinian Salinity Crisis" is indicated by an almost total desiccationf the Mediterranean and the Black Sea. From the Black Sea boreholes stromatolites and gravels are known from a depth of 864 m. They correlate with the deposition of the Arenazzolo sand in the Mediterranean region, with the formation of submarine canyons, of salt and gypsum deposits.

The *upper boundary of Pliocene* does not coincide with the internationally acccepted 1.8 m yr B.P. In the present paper the Matuyama/Gauss paleomagnetic event i.e. *ca 2.4 m yr is adopted* as the Pliocene/Pleistocene boundary in accordance with the

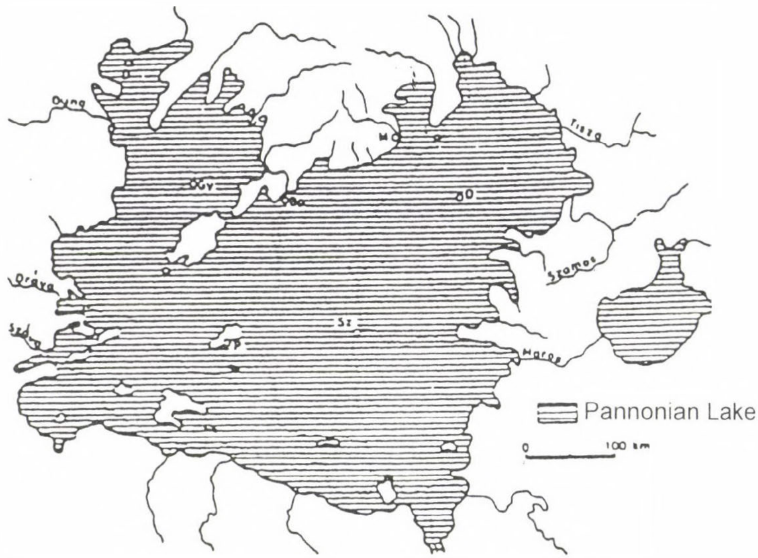


Fig. 1. The Pannonian Lake during its maximum extension (JÁMBOR, Á. 1987)

recommendation of the Hungarian Commission on Stratigraphy from 1988. The further subdivision of the Pliocene have been controversial as well. *MN zones are taken for the basis* with a tripartite Pliocene with *lower* (MN 14), *middle* (MN 15) and *upper* (MN 16) parts (KRETZOI, M. and PÉCSI, M. 1979; KORDOS L. 1992); *(tables 1 and 2).*

The role of *endogenous and exogenous impacts in relief formation* and their interrelationships have been evaluated in a different way for Hungary and abroad. Some experts give preference to tectonic movements and even more specialists are adhered to the prevalence of exogenous forces in landform evolution. There are specialists emphasizing the significance of joint effects of the two in shaping landforms. The topics studied and the aspects of geomorphological research in Hungary have been influenced by the deflation theories of L. LÓCZY (1913) and J. CHOLNOKY (1918), the concept of climatic geomorphology of B. BULLA (1954), the model geomorphic evolution by alternating erosion and accumulation proposed by M. PÉCSI (1980).

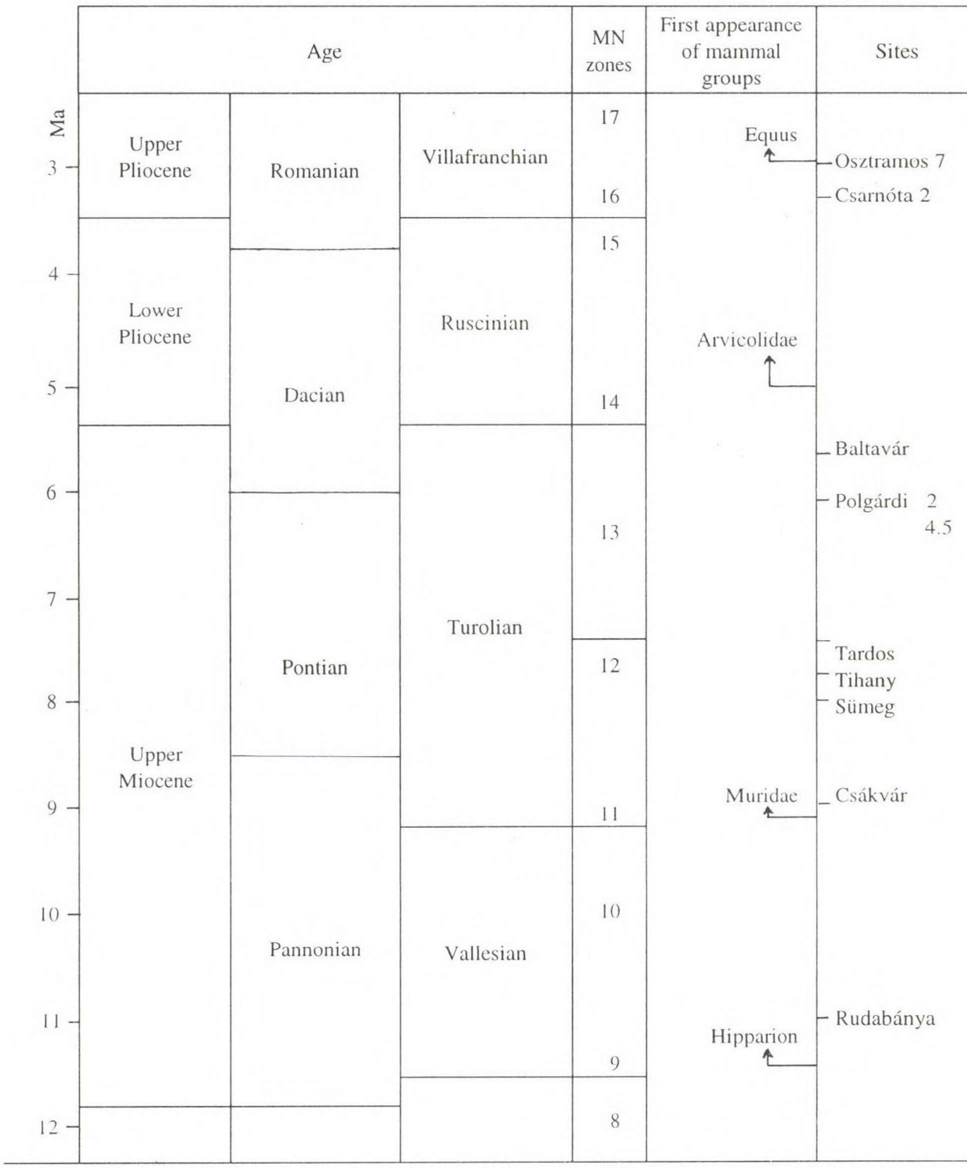
These investigations might be intriguing because the process *of elimination of the brackish inland sea and turning the area into land* can be traced so well in few regions and the Carpathian Basin belongs to these. This research includes topics of the filling up and desertification of the inner part of the Carpathian Basin, issues of pediment formation and its chronology, emergence of the river network. It is also linked to problems like the Plio/Pleistocene boundary, the age of red clays and terra rossa formations and the extent of crustal movements during the Late Cenozoic.

Table 1. Correlation of the Central and Eastern Paratethys (KRETZOI, M. 1987)

Approximate age (m yr)	Mediterranean biochronology			European terrestrial biochronology[2]								Central Paratethys	
	Codes					MN Zone-Codes						Lithostratigraphy	
												Carpathian Basin	
	Foraminifer zones	Nannoplankton zones	Stage[1]	Group	Age Stage*	POMEL 1853	GANDRY 1878	CRUSAFONT 1971	C-F-F (1972	CRUSAFONT 1974	MEIN 1975	KM[3]	RB[4]
5	N-18	NN-13	(Tabianium-Zancleum)	Barótium	Ruscinium		14	22	11	23	MN 14	Danubian	
6	N-17	NN-12	Messinium	Battavárium*(= Turolium etc.)	Bérbaltavárium*	6	13	21	10	22	MN 13		Transdanubian
7		NN-11	Tortonium (s. str.)		Hatvanium*						MN 12	Pannonian	
8					Sümegium								
9	N-16				Csákvárium						MN 11		
10		NN-10		Eppelcheimium* (=Vallesium)	Rhenohassium*		12	20b	9	21b	MN 10		Premartonian
11	N-15	NN-9	Serravallium		Bodvaium*			20a		21a	MN 9		
12					Monacium*								
13	N-14, N-13	NN-8			(Oeningium)*	5	11	19b	8	20b	MN 8	(Mediter-ranean)	(Sarma-tian)

[1]Traditional, so called mixed (bio-litho) taxa; [2]Biochronological units; of them* with lithostratigraphic significance as well; [3]Author's proposal (KRETZOI, M.–PÉCSI M. 1979), [4]Recommendation of the Pliocene Subcommission of Hungarian Commission on Stratigraphy

Table 2. Fauna of the classic site Polgárdi (N2) evidencing to the regression of the Pannonian Lake (KORMOS, T. 1911, KRETZOI, M. 1952, KORDOS, L. 1993), at least in this region of the Transdanubium. According to KORDOS (1993) vertebrate fauna at Polgárdi can be correlated with vertebrate fossils found in terrestrial deposits at Crevillente N6 site (Spain) where Messinian marine and terrestrial sediments are intercalated.



Significant warm-dry intervals during the Late Cenozoic and at the lower boundary of the Quaternary

1. At the boundary of Late Cenozoic and the beginning of the Quaternary the author identified *three significant warm and dry intervals* in the Carpathian region based on geomorphological investigations and proved by sedimentological, geochemical, paleontological, absolute chronological and paleoclimatic evidence.

In the Pannonian Basin the geomorphological levels are covered by sediments containing fauna remnants enabling conclusions about the occurrence of several intervals with warm-dry and warm-hot climates following the Sarmatian. Although based on the sea level fluctuation curve (HAQ, Z. *et al.* 1987) POGÁCSÁS, Gy. *et al.* (1987) disclosed a sedimentation gap in the Békés Basin at 10.5–11.0 m yr B.P. no evidence of that have been found in the mountains and mountain foreland (*Fig. 2*).

a) Following the post Sarmatian the *first* interval characterised with dry-warm ecological conditions (Ophisaurus, Gerbilina fossils) was the *Sümegium* (7–8 Ma, zone MN 12).

b) The *second* warm-dry interval was the *Bérbaltavárium* (counterpart of the Messinian Salinity Crisis in the Mediterranean). This was the main time of pediment formation. Typical vegetation was macchia, shrub, turning into grassland of semidesert character. Its age is 6.3–5.0 Ma, and belongs to the zone MN 13. Grey, greyish yellow sand deposits formed in considerable thicknesses, often cemented by carbonates under arid climate. In places covered with shallow water, owing to the short route of transportation, unsorted clayey-sandy formations, 'variegated clays' emerged. According to the latest data sand surfaces and desert crusts formed, subsequently covered by red clay and terra rossa in the late Ruscinium and Csarnotanum and Quaternary sediments.

c) The *third* important warm-dry interval was the *Villányium*. At that time lower, poorly developed pediment surfaces formed; alluvial fans (e.g. Kislángium with camel and ostrich), very probably the most ancient, so called 'warm loesses' (Dunaalmás, Szekszárd) and terra rossa desiccation crack infillings with carbonate veins and concretions (Dunaalmás, Villány) belong here (3.0–2.0 Ma, zone MN 16–17).

2. Author attempted at the interpretation and explanation of the *evolution of plains in semidesert and steppe environments which has been a topic for a century-long debate.*

A surface evolution hypothesis on the so called desert phase of the Carpathian Basin have been discussed by many (also from geological, geomorphological and paleontological aspects). A limited validity of a desert phase during the Pannonian and Pontian, voiced by LÓCZY, L. (1890, 1913), was accepted by several scientists but in the Pleistocene exclusively and with reservations. Among others, BULLA, B (1953) adhered to a concept according to which during the whole Tertiary (including the Pliocene) tropical, subtropical climates prevailed and tropical weathering at that time had promoted peneplanation (between the 1940's and 60's the Plio/Pleistocene boundary was drawn between 600 thousand and 1 million yr B.P.). Since BULLA, B. (1947, 1954)

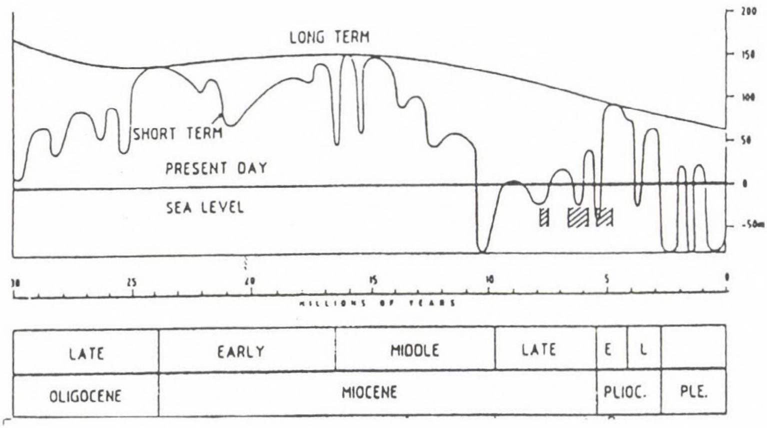


Fig. 2. Eustatic sea level oscillations during the past 30 m yr of geohistory and their relative extent (after HAQ *et al.* 1987). Comparing global sea minima (5.2; 6.3; 7.8; 10.4 Ma B.P.) with hiatuses identified on the northern shelf of the Pannonian Basin (5.4–4.6; 6.8–5.7; 7.9–7.6 Ma B.P.) and with the age of the hiatus represented by the regional discordance surface separating the synrift and postrift sediments with changing duration from place to place (10.5 m yr B.P.) there is a remarkably close correlation

unambiguously excluded the existence of a deflation desert climate at the end of the Pannonian-Pontian, this scientific problem was hardly touched upon later.

a) In arid and semiarid regions *crusts of 2–5 cm thickness* are typical which form on various geomorphological levels. Samples were *collected by the author from several localities* in the Carpathian Basin for the last 15 years (Mogyoród, Bana, Bábolna). Recently post-graduate research students under his guidance have proceeded with assembling convincingevidence from parts of Transdanubia that first became land during the Upper Miocene, e.g. from several places of the Tapolca Basin, Kemeneshát, southern foreland of the Kőszeg Mountains, and from mountains and forelands formed during the Bérbaltavárium (FÁBIÁN, Sz., KOVÁCS, J. and VARGA, G. 1999). On the geomorphological surfaces in semiarid environment sand deposits, sediments deposited by torrents cutting into deltas and alluvial fans accumulated in a considerable thickness, siliceous desert crusts formed, pans of evaporites appeared over the Pannonian clay or between variegated clays in thicknesses of 0.5–1.5 m. Another evidence are *gravels with desert varnish*, ther global appearance of which is confined to semiarid-arid environment and they were also found in several places in Hungary in gravel beds of Pliocene alluvial fans and fluvial accumulations (Mogyoród, Csömör, Bábolna, Bratislava, Bazin, Vienna Basin etc.).

Wind-polished, reddish brown coloured concretions – on the basis of macroscopic features (colour, shape, surface) – show strong similarities to desert pans; they are of flattened, irregular or oval shape, with 2–10 cm diameter and 0.5–2.5 cm thickness. Also they are akin to the latter as far as their mineralogical and chemical composition and textural characteristics are concerned. Comparative studies between the desert crusts of the Sahara and that of the Carpathian Basin – based on thermal analysis (BERECZ I. *et al.* 1983) – established that both of them are composed of amorphous silica (SiO_2xH_2O), a large amount of cryptocrystalline goethite (FeOOH) and of a small amount of $CaCO_3$ (*Fig. 3*). Studies under polarising microscope and using X-ray analyser connected to SEM have indicated a genetic similarity between the two samples collected from distant localities. Ferrous, manganese, siliceous varnishes and pans formed through the evaporation of intermittent lakes and of sediment moisture in deflation hollows. Apart from their typical main elements they contain several trace elements (K, S, Cl, P) referring to dissolution–precipitation diagenetic origin. This process is probably of biogenic origin, with the precipitation also promoted by algae.

An other important observation is (JUX, U. 1983) that these hardpans precipitate in clayey-sandy deposits overlying sand sediments with ascending alkali pore waters and in their surroundings limestone-dolomite and gypsum formations are found. Siliceous desert pans are characteristic in areas with less than 130 mm per year annual precipitation and 16–24 °C annual mean temperature but due to their inheritance (in the Sahara they never occur in younger levels) it is difficult to give precise definitions.

This is why the concretions at Mogyoród are supposed by the author to be products of a similar *process of evaporation*. Their similarity with the samples from Algeria and indications in their environment (enrichment in boron at the boundary of the fluviolacustric sequence and the underlying sand at Mogyoród, black manganese cover containing barium, silica crusts and gravels with desert varnish, root pseudomorphoses in the sand deposit) all are in support of the above concept.

b) A *joint evaluation of landforms* (e.g. pediment surfaces, debris and alluvial fans), the *presence of gravels with desert varnish and desert crusts, the origin of the basaltic residual hills by deflation, together with the succession of the vertebrate fauna and the impact of the salinity crisis on the Carpathian Basin,* author proposes a high probability of *correlation of the Messinium with the Bérbaltavárium* (SCHWEITZER, F. 1992, 1994, 1997). This period might coincide with an almost total desiccation of the Mediterranean Sea ("Messinian Salinity Crisis") when salt and gypsum deposited in its basin.

Data referring to warm-dry and hot-dry phases could be found in the borehole profile Jászladány N1 with a total extension of 950 m, at a depth of 432–720 m (RÓNAI, A. 1985, *Table 3*). In this profile the climate of the warm deciduous forest also indicated by a richness of species (Upper Pannonian substage, between 930–740 m) definitely differs from the semidesert climate of the Levantan substage with bleak, unforested vegetation representing a stratigraphic marker.

A warm Mediterranean phase of the formation of terra rossa: the Csarnotanum

Geomorphological positions of the typical red clays and reddish clays, their geochemical analyses and conclusions drawn. As far as the dating of desert crusts is concerned, red clays, reddish clays, soil horizons and their geomorphological position

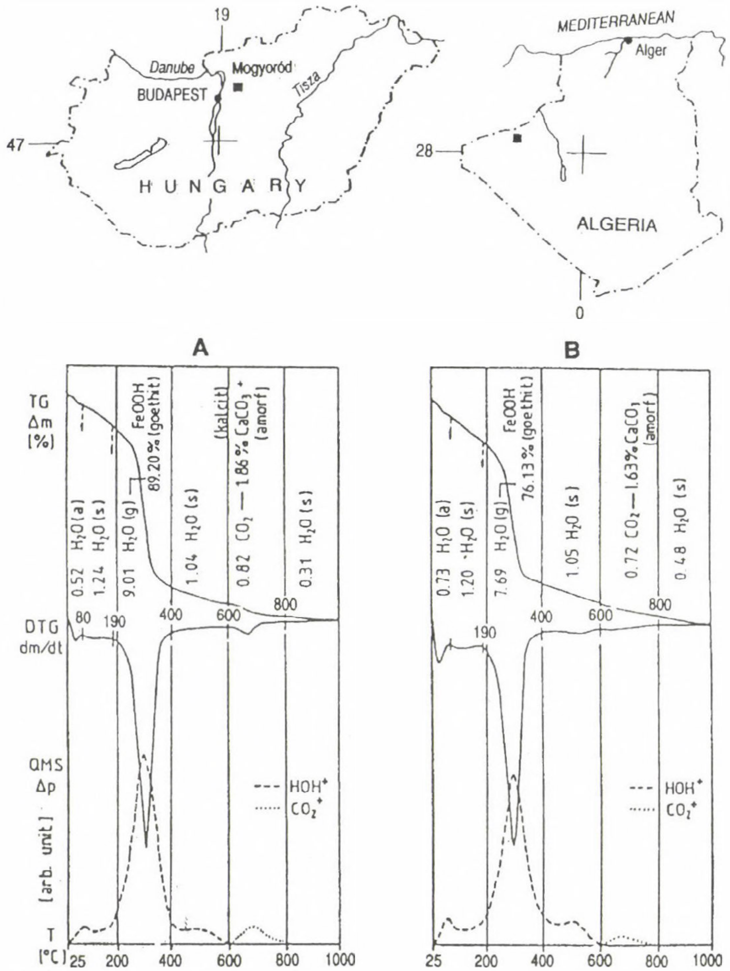


Fig. 3. Results of thermal analyses of carbonaceous–ferrous–siliceous concretions found in a thick sand deposit coverting delta sediments at Mogyoród, Hungary (A) and in Algeria (B)

might be instructive. There are several concepts on the position, spatial distribution, characteristic features of red clay and of red coloured clayey formations both in the international literature (BÜDEL, J. 1950, KUBIENA, W.L. 1956, KUKLA, G. 1987) and in Hungarian sources (STEFANOVITS, P. 1958, BIDLÓ, G. 1982, BORSY, Z. and SZÖŐR, Gy. 1981, PÉCSI, M. 1985, KRETZOI, M. *et al.* 1982). In the opinion of some experts, formation of red clays is associated with bauxite formation, however, most of the specialists share Kubiena's (1956) position attributing red clays to two different processes:

Table 3. Climatic cycles according to the pollen spectra analysis of the borehole Jászladány N1 (RÓNAI, A. 1985)

Age	Depth of sample	Climate
Holocene		moderate dry
Q_4	0–6	
↓	↓	↓
Lower part of the Pleistocene		
Q_{1-7}	285–303	warm–humid
Q_{1-6}	303–333	moderate–arid
Q_{1-5}	333–347	warm–humid
Q_{1-4}	347–363	warm and moderate–arid
	363–366	cool-dry spell in the wake of the stage
Q_{1-3}	366–397	warm–humid
Q_{1-2}	397–410	moderate–humid
Q_{1-1}	410–432	warm–humid
Upper member of Levantan	432–550	warm–dry
Pl_{3-2}		
Lower member of Levantan	550–740	hot–dry
Pl_{3-1}		
Last phases of the Upper Pannonian		
Pl_{2-3}	740–800	warm–moderately humid
Pl_{2-2}	800–860	warm–dry
Pl_{2-1}	860–930	warm–very humid

1. Red clays form in humid or dry conditions under the impact of rubefication requiring warming effect;

2. They are the result of lateritisation taking place under permanent humid and warm conditions and leading to bauxite formation.

– *On the one hand* – from the aspects of landform evolution – red clays should be considered *"correlative sediments"* providing clues for the reconstruction of past geographic conditions (paleoclimates, soils, and relief-forming processes such as erosion or tectonic movements). They are morphoclimazonal phenopmena and were confined to warm subhumid grove environments on active pediments or on those in the stage of dissection with the prevalence of clayey weathering. In situ formed red clays could easily erode (partly or wholly) owing to increased slope inclination or for other reasons; on gentle slopes it had been accumulating or intercalated by other sediments, whereas in smaller sediment traps soil series with red clays might be formed. Therefore, red clays, as "correlative sediments" should be regarded as indicators of the "climax of pediment conservation".

– *On the other hand* geomorphological and lithostratigraphic position of *morphoclimazonal red clays* represents a time interval of specific geomorphic conditions. In such a sense loess series can be likewise considered a sequence having formed under semiarid conditions of periglacial zones with forest steppe and steppe vegetation and on the steppe margins of certain deserts during cold phases of the Pleistocene.

Lithostratigraphically red clay formation is underlying the most ancient loesses and loess-like sequences in the Middle Danube Basin, in the southern zone of the Russian Plain and in China. In the ancient loesses there are frequent intercalations of soils of reddish ochre colour, however, none of them are red clays but steppe soils formed under warm-dry, subhumid climates (chestnut or cinnamon soils). These soils are especially typical of the lower Pleistocene loesses on the Loess Plateau in China (e.g. Baoji) and in Central Asia (Tajikistan, Chashmanigar), where ca 20 of these soils are superimposing each other almost directly or with thin loess intercalations making up total thicknesses of 50–80 m. These reddish soils – owing to a low degree of clayey weathering – are paleoecologically different from red clays, and are alternating with thick loess pockets having formed during the Early Pleistocene. Soils represent the moderately subhumid, warm semiarid steppe zone, whereas the interbedded so called warm loesses might have formed under slightly more humid semiarid steppe climates. The time interval of the formation of the latter is between 2.4–1.7 m yr B.P. 1–2 m thick red clay units overlying each other form sequences of more than 30 m thickness. These had formed between >5–2.4 m yr B.P. In the Middle Danube Basin red clays superimpose the Upper Miocene or Upper Pannonian sequence or more ancient formations.

Its type localities can be studied in several exposures (Gödöllő Hills /Bag/, Hatvan brickyard, Mogyoród, Gyöngyösvisonta /Rókus Hill/, Kulcs, Dunaföldvár, Szekszárd and Bátaszék brickyards etc.) in Hungary. In boreholes (Pécs-Postavölgy, Szekszárd, Dunakömlőd, Dunaföldvár, Dunaszekcső, Tass, Tengelic etc.) typical red clays are underlain by Upper Pannonian formations (with bentonite interbedding), whereas the sandy-silty series above them are covered by old loesses. Lithostratigraphically domestic red clays are situated between the Bérbaltavár sand and variegated clays or reddish soils (terra rossa) having developed after the formation of bentonite at Gyöngyösvisonta and found under the oldest loesses. Geomorphologically their formation had followed the development of older pediment surfaces and early Pannonian abrasion terraces and lasted up to the appearance of younger Late Pliocene–Early Pleistocene (Villányium) foothills, in the correlative sediment of which only the debris or redeposited material of red clays is present as semipedolite. Chronostratigraphically this period – similar to the Chinese red clays – falls in between 4.5–2.5 m yr.

The profile of the Dévaványa core drilling (RÓNAI, A. 1985) was revisited and studied. 10 red clay horizons could be identified between 1000–700m. According to paleomagnetic datings they might have formed between epoch 5 and the Gauss–Gilbert boundary (>5.0–3.3 m yr).

In summary, *the formation of red clays comprise a paleogeographic interval of similar duration as loess formation. Geomorphic evolution and sedimentation, however, had taken place under completely fferent climate morphological conditions. The alternation of loesses and paleosols might be explained by climatic cycles when loess repre-*

sents the cold phases, while soils the warm ones. In contrast, *red clays are the remnants of warm and humid (subtropical) climate phases. During the warm semiarid interruptions no substantial sedimentation occurred. The material was partly washed away even during the interval of red clay formation , partly contributed to the mineral mass of subsequently accumulated red clays.* In situ *red clays witness to climatic fluctuations between warm-humid (subtropical) and warm-semiarid cycles and are found typically in the middle and lower sections of pediment surfaces.*

Weathering products of warm-humid (subtropical) climate are kaolinite-halloysite, whereas under moderately warm-humid and semiarid climates illite-montmorillonite formed and variable carbonate paragenesis took place. These two different products (red clays and reddish clays) were separated by SZÖŐR, Gy. (1993) using several geochemical parameters. The changes in the total amount of uranium and thorium (U_{ekv}) and percentage iron dioxide (Fe_2O_3) is a good example (*Fig. 4*). The explanation of this regularity refers to the weathering-solving processes controlling mineral paragenesis.

Climatic cycles induced by oscillations of Earth orbit parameters used as explanation for Pleistocene cycles by M. MILANKOVIČ (23 ka, 100 ka, 400 ka) obviously were valid during the Pliocene and Miocene as well.

Semiarid climatic conditions and the main period of pedimentation

Pediment surfaces (basically different from those developed under Pleistocene periglacial conditions formed in the Carpathian Basin *under semiarid climates.*

The typical pediment surfaces – except for cryopediments and cryoglacis – might have formed in *three phases and during three different stages.*

a) *Sümegium* (7.5–7.0 Ma B.P.), MN 12 zone. The beginning of pediment formation. Under warm-dry climate geomorphological levels formed at 400–420 m a.s.l. in the Buda Hills (Széchenyi Hill) and at 300–350 m a.s.l. in the Gerecse Mountains. This is corroborated by the fauna of the type locality (e.g. *Ophisaurus* and *Gerbillina*) with expressed thermoxerotic ecological requirements.

b) *Bérbaltavárium* was the onset of formation of typical pediment surfaces (6.3–5.0 Ma B.P.), MN 13 zone. Landscapes without woodland, with arid shrubs, sometimes grassy, turning into semideserts. The development of pediments can be attributed to processes of areal erosion under semiarid climates with a simultaneous uplift in the surroundings of the Carpathians. The arid semidesert conditions, deficit of moisture are indicated by the absence of travertine horizons. For instance, along the eastern margin of the Buda Hills pediment surfaces were formed between 370–230 m a.s.l. divided into two levels by a travertine horizon at 275 m indicating the renewal of karstic springs: an older one between 370–270 m, of Bérbaltavár age and a younger one (270–230 m) of Villányium age (*Fig. 5*).

c) Villányium (3.0–2.0 Ma B.P.), MN 16–17 zone, along with the lowering of pediment surfaces formed during the Bérbaltavárium there was a process of formation of

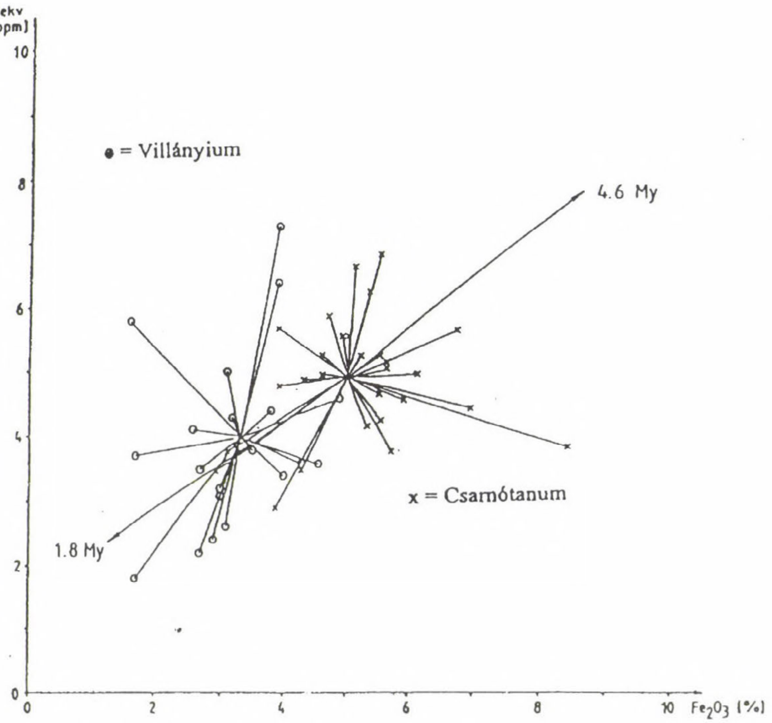


Fig. 4. Separation of the typical Pliocene red clays from the Lower Pleistocene paleosols and sediments based on the U-Th and Fe_2O_3 content (SZÖŐR, Gy. 1993)

atypical pediments. Derasional valleys (filled up by reddish, purplish soils) were cut into their sloping surfaces. Pediments extended over terraced alluvial fans.

The problem of the Tertiary–Quaternary boundary and other possible boundaries

It was even earlier that opininons were strongly divided among the experts of Late Neogene and Quaternary concerning the Neogene/Quaternary boundary. The International Geological Congress (London, 1948) made recommendations that the Pliocene/Pleistocene boundary should be drawn at the bottom of the Calabrian layers, the first marine sediments containing cold tolerating foraminiferans. Later this was determined by Arias, C. *et al.* (1980) using paleomagnetic dating at ca 2 Ma B.P.

The Neogene–Quaternary boundary – on the basis of the climatic calendar by MILANKOVIČ, M. (1930) and BACSÁK, Gy. (1942) and the phases of Alpine glaciation by Penck–Brückner – earlier was established at ca 600 ka B.P. which coincided with the first significant glaciation in the Alps. After there had been found traces of several previous glacial stages (Donau /Eburon/, Biber /Praeteigelen/), the duration of the Pleistocene had been extended, first up to 1.8 Ma, then (by others) to 2.4 and 3.0.

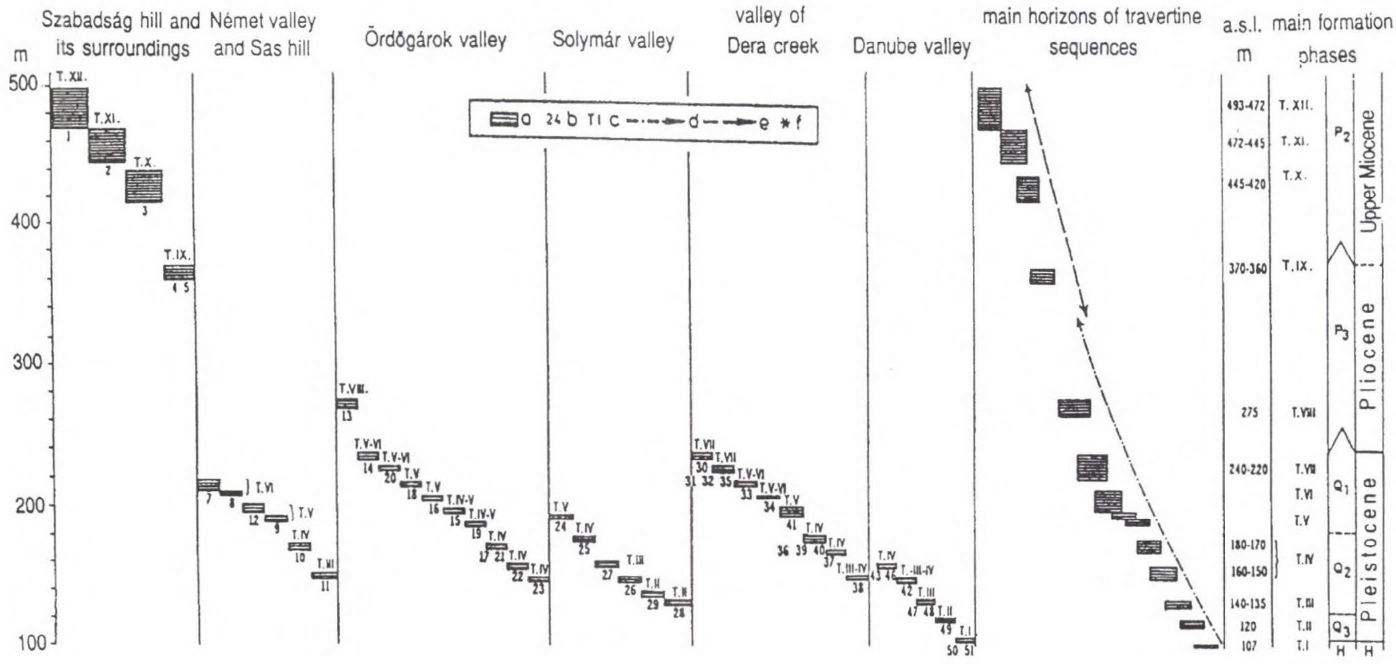


Fig. 5. Relationship between the Pannonian abrasional terraces (travertine horizons TXII–TIX), pediment surfaces (H1–H3) and travertines confined to the valleys of the Buda Hills (TVIII–TI) (SCHEUER, Gy and SCHWEITZER, F. 1972). – a = travertine horizons; b = occurrences; c = TI–TXII – main phases of travertine formation; d = beginning of the formation of erosional valleys and travertines confining to them; e = stadial tectonic movement of the János Hill and Szabadság Hill with a dominant trend of uplift; f = travertines confining to tectonic movements

This trend was corroborated by the fact that the wake of the Quaternary has always been parallelled with the appearance of the Early Man. This is confirmed by the findings at Diring, Central Siberia at the same time raising the problem of non-tropical origin of man. The age of the Olduvai site is 1.8–1.7 m yr, and that of Coobi Fora (eastern shore of Turkana Lake) is 2.2–2.0 m yr. The cultural layer of the Diring site is estimated by MOCSANOV, I.A. (1983) as 3.2–2.2 m yr. The lower limit of the cultural layer N5 is 3.2–3.1 m yr old (according to the paleomagnetic datings it is 4.2–3.9 Ma) while the age of the upper limit is 2.4–1.8 Ma. Findings of this latter, younger paleolithic site (in Mocsanov's opininon) shows remarkable similarity to those of Olduvai. By several Russian experts the Early Man of Diring must have lived in a very cold environment. The author of the present paper considers finds at Diring much older than those of Olduvai. The findings bearing sediments were deposited by torrents: sorted sand, gravels with desert varnish (even stone tools are covered with varnish), ventifacts. This sequence in some places is covered by red clays and sand and (Eopleistocene) sediments with intercalated paleosols.

These observations might induce a new approach to the paleoclimatic interpretation of the paleogeographic conditions during the Pliocene epoch.

It is conspicuous that the sand series of Diring with findings (61° 12' N latitude, 128° 28' E longitude) containing gravels with desert varnish (serir) and overlain by red clays are very similar to those met by the author in Alaska, USA (63° 54' northern latitude, 145° 11' western longitude) and in Yukon, Canada (63° 56' northern latitude, 138° 30' western longitude) but without paleolithic findings.

In 1984 the International Commission on Stratigraphy (with a compromise solution primarily based on the Olduvai Bed layers) *drew the Neogene-Quaternary boundary at 1.8 Ma B.P.* This boundary, however, on the basis of geological and biostratigraphic evidence hardly could be linked to any important geohistorical event.

Comparative paleontological investigations carried out earlier by JÁNOSSY, D. (1979), KRETZOI, M. (1962, 1969) and more recently by KORDOS, L. (1987) have proven an enrichment of fauna succession starting with 1.6–1.5 Ma B.P. and reaching its climax at 1.2 Ma, but deteriorating soon (by 1 Ma). Based on this KORDOS, L. (1987) suggested an event in terrestrial biochronology to be correlated with 1.7–1.6 Ma obtained by the studies of marine sediment formations (Olduvai event). Paleontological research has shown that this interval marks a turning point in the development of mammal fauna in the Carpathian Basin. Species numbers began to grow at 1.6–1.5 Ma B.P. which largely corresponds to the lower boundary of the Biharium, while its upper limit is put unambiguously to the Brunhes/Matuyama boundary, i.e. 0.72 Ma (BERGGREN, W.A. *et al.* 1985). The lowermost part of old loess series named Paks loess (below PD paleosols) can be correlated with the Lower Biharium as well.

Chronologically the older (V/a) member of the Danube terrace V (covered by a younger freshwater limestone than the travertine of Kislángium age) also belong here. Glacis formation in larger valleys and intramontane basins and dissection of the youngest pediments into interfluvial valleys are confined to this time interval.

In Hungary the Plio-Pleistocene boundary is drawn along the Matuyama–Gauss paleomagnetic boundary, i.e. at 2.4 Ma. At thet time significant tectonic–paleogeographic–environmental changes occurred in the Carpathian Basin.

By this time dry-warm steppe fauna of the previous Villafranchium (3.5–2.5 Ma B.P.) had become impoverished (JÁNOSSY, D. 1979, KRETZOI, M. 1952, 1969, KORDOS, L. 1988), then disappeared and between 2.4–2.0 Ma B.P. a new faunistic event occurred with the growth of the number of species.Very probably this event makes 2.4 Ma suitable to draw Plio–Pleistocene boundary in the Carpathian Basin. According to RÓNAI, A. (1972, 1985) deep boreholes in the Great Hungarian Plain show a general trend of ther deterioration of the climate, but with interruptions. Beside uneven cooling further fluctuations are created by drier or moister phases. During the period between 2.4–1.8 Ma five climatic intervals could be identified on the basis of the material of boreholes.

There are experts, however, insisting on the lower boundary of the Pleistocene to be drawn along the stage 22 of the oxigen isotope scale. This corresponds to the cold peak of the Pleistocene dated at 0.8 Ma B.P. In the Russian literature this is the Pleistocene–Eopleistocene boundary.

Author' research has contributed to the solution of the problem of the oldest "warm loesses" formed at the boundary between the Pliocene and Pleistocene.

The formation below the loess series in non-subsiding hill areas or basin margins as a rule contains hiatuses. In some geomorphological positions, however, where in between terra rossa (reddish soils) or superimposing them and, in some places enclosed by travertines, loesses with loess dolls and loess-like deposits have survived (Szekszárd, Bátaszék, Dunaalmás). Found by the author and studied by D. JÁNOSSY fauna association (at Dunaalmás) was situated in the terra rossa underlying loess and loess-like sediments and belonged to the Upper Villányium. These loesses are pale pink, not too thick and do not contain fauna. The oldest loesses of Tajikistan and China are of similar age, but in Alaska even more ancient loesses (3 m yr) can be found.

Studies by author based on investigations into pediment levels, correlative sediments of various age formed on their surface, the oldest terraces and travertine horizons, contributed to the establishment of the extent of crustal movements during the Upper Miocene and Pliocene. So, the Buda Hills uplifted 130 m during the Upper Miocene, 150–160 m in the Pliocene and 120–130 m in the Pleistocene.

Tectonic movements had shown spatial differentiation during the Pliocene. They amounted to 100–120 m in the Mecsek Mountains, 50–60 m in the Gerecse and 40–50 m in the Bakony.

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COMPARATIVE STUDIES OF TERRACES AND LOESS SECTIONS IN THE VICINITY OF KRASNOYARSK IN THE YENISEY VALLEY

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MANFRED FRECHEN[3]

Introduction

For several decades a recurring topic of investigations – both from theoretical and methodical aspects – have been comparative studies of characteristic valley sections of the European and Asian big rivers. An essential part of these studies have been joint field surveys with the involvement of local experts of international experience focussing on the geological evolution of fluvial terraces and loess profiles.[4]

In the beginning of the 1990's at a conference of Quaternary geology and geomorphology held in Central Yakutia (centred in Yakutsk) Hungarian experts reported on the loess profiles and chronological subdivision of the Central Danube Basin. Previously a similar lecture was held at an INQUA Loess Commission meeting organised on the Chinese Loess Plateau, at Xian and Luochuan (PÉCSI, M. 1987a). An agreement had been reached that specialists involved in the study of Quaternary fluvial sediments and loess-paleosol sequences in China (Xian, Beijing, Nanjing) and in Siberia (Novosibirsk, Krasnoyarsk) should study loess and terrace formations of the Danubian Basin, while Hungarian experts would be given a chance to get acquainted with similar localities along the major rivers of China and Siberia in the framework of exchange of methodologies and experience.

Of these field excursions the ones covering the Middle Yenisey Valley took place in 1992 and 1995 organised by A.F. YAMSKIKH, head of the Laboratory of Pa-

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[4] During field excursions of international conferences or in the course of bi- or multilateral research projects Hungarian specialists had the opportunity to participate at field excursions along the representative stretches of large rivers of Central and Western Europe (Danube, Vistula, Elba, Rhine, Main, Seine...) then those of Eastern Europe and Asia (Dniester, Don, Volga, Sirdarya, Amudarya, Yenisey, Lena, Aldan, Hwang, Ganges...) and of the Mississippi-Missouri (USA) and Paraná-Paraguay (Argentine)

leogeography, Krasnoyarsk Teachers's Training School, who had elaborated loess series of the river valleys along Yenisey and tributaries in his dissertation of doctor of sciences (YAMSKIKH A. 1992). The Hungarian party was represented by M. PÉCSI and F. SCHWEITZER, Quaternary and loess experts from the Geographical Research Institute Hungarian Academy of Sciences (GRI HAS). Relevant type horizons of sections containing terrace material and loess sediments were studied jointly and sampled for subsequent TL analyses. Dr. M. FRECHEN from the Institute of Geology, University of Cologne (Germany) undertook the task of dating. Later Dr. FRECHEN at the invitation of the Siberian partner institute studied the key sections personally and collected samples for further TL analyses in 1995.

Two paleogeographers from the Krasnoyarsk institute came to Hungary in autumn 1992 and made study trips to examine key loess profiles in more detail. The first part of the analyses was finished by 1997. First and preliminary summary of ideas and results is to be attempted below.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 014152.

Discussion

On the Yenisey a huge dam was constructed at Divnogorsk, raising water level by 100 m, before the river leaves its mountain reach. The medium level of the dammed section (up to Abakan) is situated at 243 m a.s.l. while below the reservoir (down to Krasnoyarsk located at a 40 km distance) the medium level changes between 142 and 130 m. This way along the (300 km long) water reservoir[5] valley terraces lower than 100 m have been inundated. Natural fluvial terraces of Yenisey might be studied recently in the Krasnoyarsk Basin and along the northern foothills of the Eastern Sayan Mountains (YAMSKIKH 1992).

Terraces along the middle and upper reaches of Yenisey (according to the variation of mountain and basin morphostructures) differ in their relative altitudes and also their number is changing. A similar pattern can be recognised along the valley sections of the Danube breaking through the Carpathians and crossing the enclosed basins (PÉCSI 1959, 1971).

Several researchers studied the geological-geomorphological position, composition and age of the terraces within the mentioned sections. These investigations and results achieved by other experts were summarised recently by YAMSKIKH, A.F. (1993) (*Fig. 1* and *Table 1*).

[5] Between Minusinsk and Divnogorsk

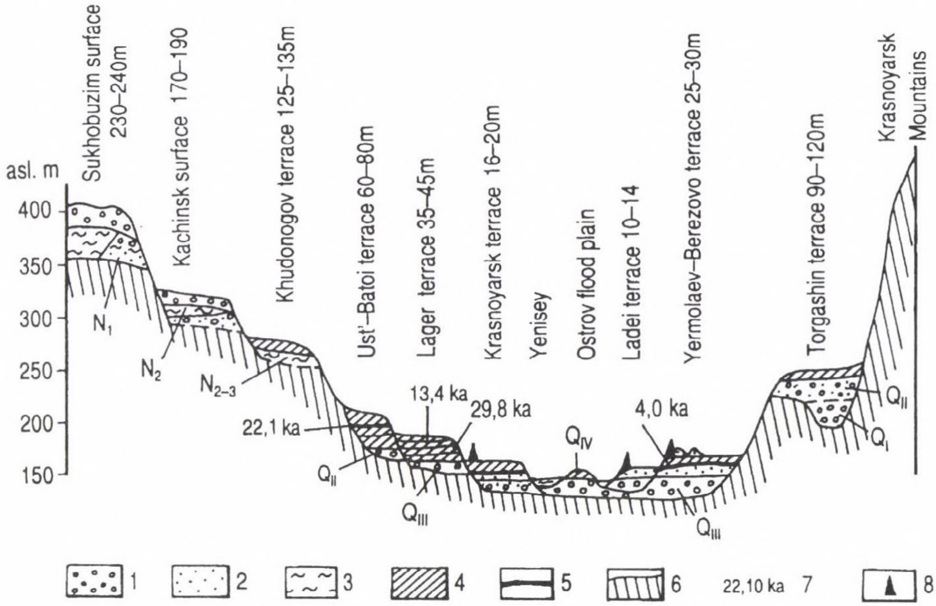


Fig. 1. Terraces of the Yenisey in the vicinity of Krasnoyarsk, ca 56° northern latitude (according to YAMSKIKH 1993). – 1 = gravel; 2 = sand; 3 = clay, loam; 4 = loess-like sediments; 5 = paleosols; 6 = bedrock; 7 = ^{14}C age; 8 = site of Early man

Table 1. Terraces along the middle reaches of Yenisey: a comparison between the different concepts of authors (After YAMSKIKH 1993)

Geomorphic levels	YAMSKIKH 1993	FINAROV 1964	ARKHIPOV 1966	BORISOV 1984
Flood plain level	4–7/12 m	–	–	2–6 m
Terrace I	5–12 m	8–14 m	10–12/8–11 m	4–8 m
Terrace II	14–18 m	15–25 m	15–18/12–15 m	12–15 m
Terrace III	24–30 m	30–36 m	23–27/17–25 m	15–25 m
Terrace IV	35–45 m	40–60 m	30–35 m	25–35 m
Terrace V	45–55–60 m	60–80 m	40–45/40–50 m	35–60 m
Terrace VI	60–80 m	100–120 m	60–65/70–80 m	60–80 m
Terrace VII	80–120 m	130–140 m	70–80/90–100 m	80–120 m
Terrace VIII	130–160 m	160–180 m	100–120/110–120 m	120–130 m
Terrace IX	–	–	–	–
Kachinsk surface of planation	170–190 m	200–240 m	130–140/150 m	–
Sukhobuzim surface of planation	230–240 m	–	–	–
Studied area in the Yenisey Valley	Krasnoyarsk Basin periphery	Minusinsk Basin (Rakovets /1969/ came to similar conclusions)	From Krasnoyarsk to the tributary of the Angara	Minusinsk Basin in the foreland of the reservoir

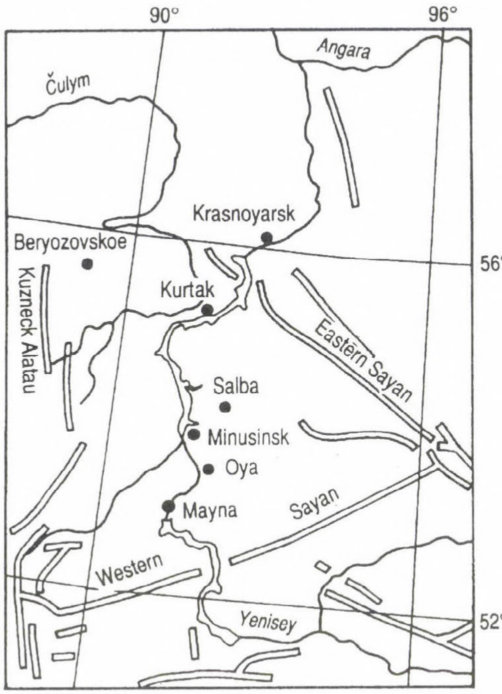


Fig. 2. Geographical setting of the Yenisey Valley

Laboratory of Paleogeography of Krasnoyarsk Teachers' Training School predominantly has dealt with the complex task of the explanation and dating of polycyclic and polygenetic (fluvial, deluvial, eolian) evolution of lower terraces in the Yenisey valley. That is why joint field surveys were focused on the geological and geomorphic position and phenomena typical of young terrace sediments and of the superimposing loess, alluvial material and slope deposits when analysing representative profiles along the river section between Krasnoyarsk and Minusinsk (*Fig. 2*).

Along with young terrace exposures young loess series (Sisim) overlying some older terraces (N°V–VI) and remnants of older degraded loess and young sand and loesses superimposing eroded loess pillars were observed, too.

Earlier investigations mentioned that in the Middle Yenisey Valley not only Quaternary terrace formations but ancient weathered rocks and Tertiary deposits could be found locally and their traces were encountered during joint field excursions. This means that some valley sections could develop well before the Quaternary, then they were partially buried and later exhumed.

Study of some representative profiles of low terraces and flood plains

The Krasnoyarsk team has been involved in terrace investigations, based on complex stratigraphic analysis of profiles of the high flood plain and those of the superimposing low terraces overlying them and also on ^{14}C datings of humus and charcoal. As a result conclusions have been drawn concerning the age of flood plains and terraces (*Figs 3* through *5*).

Shaped by an extreme (seasonal and periodical) hydrological regime normal alluvial sequences were sedimented, while in some cases, e.g. during disastrous floods, lacustric-fluvial series were formed within the dammed valley sections. These sediments

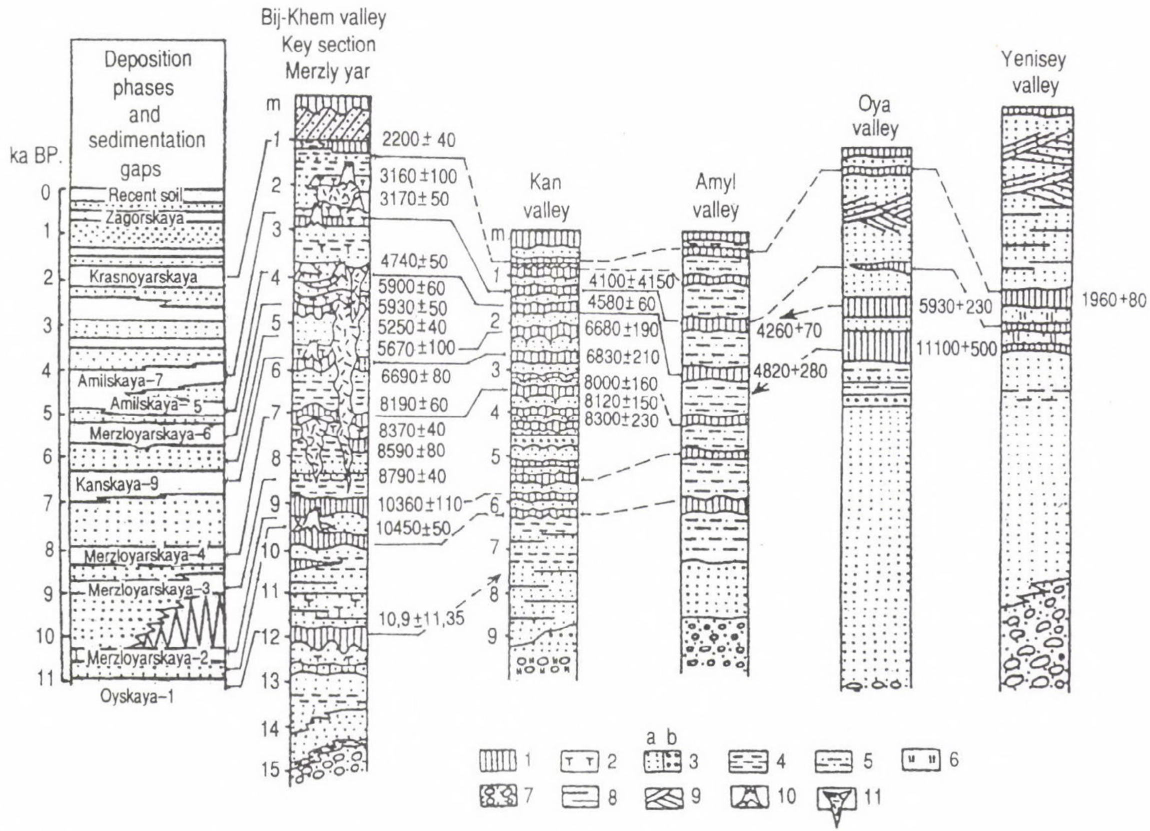


Fig. 3. Correlation of polycyclic Holocene sediments on the southern part of the Yenisey Valley. 1 = soil; 2 = peat; 3 = sand (a), gravelly sand (b); 4 = silt; 5 = loam; 6 = dusty sediment; 7 = gravel bed; 8 = horizontal layers; 9 = oblique and diagonal layers; 10 = remnants of buried forest; 11 = ice wedges

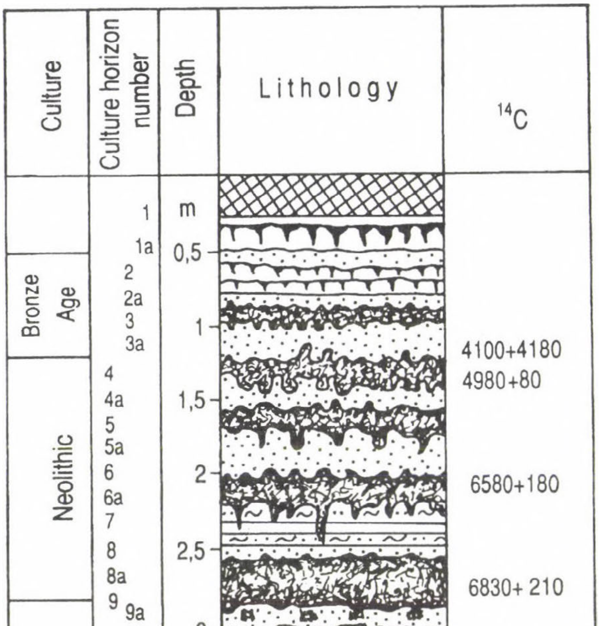


Fig. 4. Profile of polycyclic Holocene terrace in the area of multi-layered site Kazachka (on Kan River). – 1 = artificial filling; 2 = fragments of young alluvial soils; 3 = alluvial paleosols; 4 = sorted sand; 5 = non-sorted sand; 6 = loams; 7 = humous sand

are much thicker and are composed by specifically layered alluvial and basin sediments. In this way periglacial and intra-continental water regimes in Southern Siberia resulted in polygenetic and polycyclic terraces. According to YAMSKIKH (1983) cyclic climatic change led to sedimentation phases of 21–22 ka, 7–8 ka duration in the late Pleistocene, while shorter spells (400–500 years) were characteristic for the Holocene. These fluctuations are thought to be supported by radiocarbon and palynological evidence.

For the verification of sedimentological investigations of loess-paleosol sequences and fluvial sediments in the vicinity of Krasnoyarsk summarised by YAMSKIKH (1992, 1993) a joint study of the profile of the so called lower Lagernaya (Tatyshev) terrace with the overlying loess-paleosol sequence (located in Krasnoyarsk city, at October Bridge) was carried out (*Fig. 6*, Tatyshev profile surveyed by PÉCSI, SCHWEITZER and YAMSKIKH). Previously ^{14}C datings based on humus content of three soil horizons (h_1, S_1 and S_3) were performed. At a depth of 3 m the age of h_1 proved to be 13,400±70 yr, at 9 m S_1 was 22,100±80 yr and at 15 m S_3 gave an age of 29,800±2000 yr B.P.

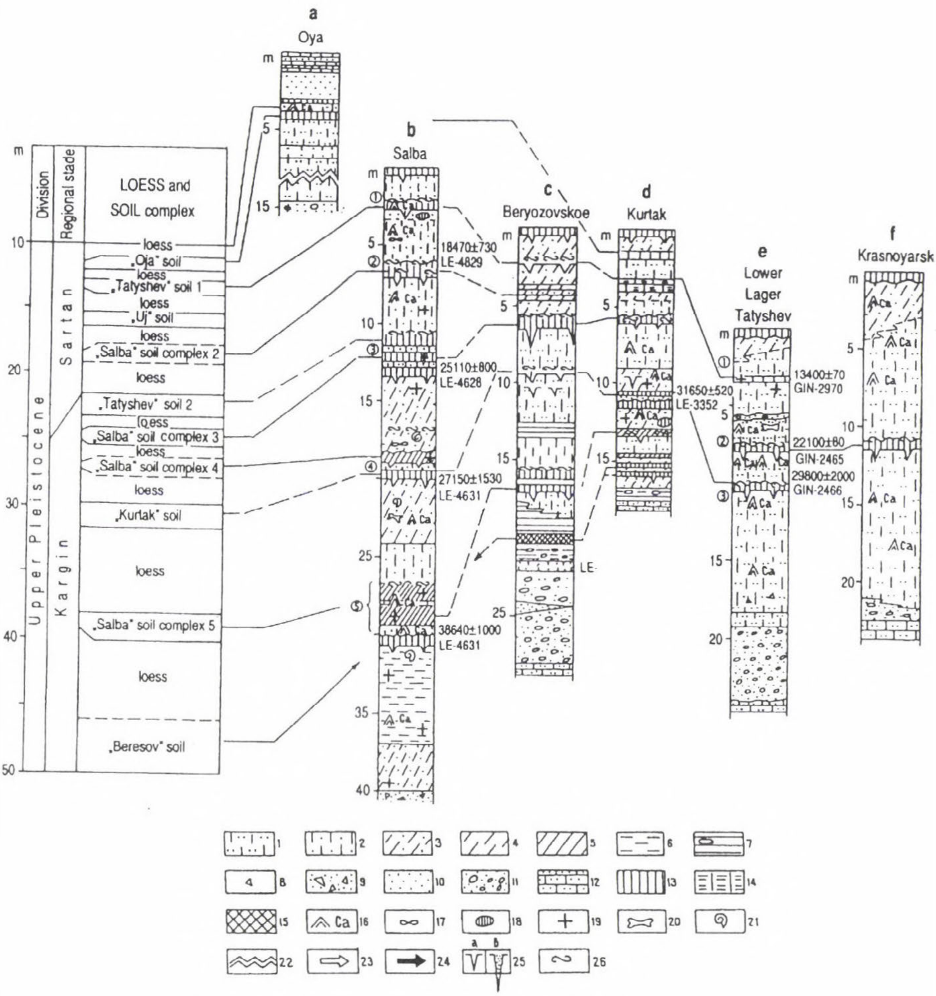


Fig. 5. Correlation of loess-paleosol profiles in the middle Yenisey Valley (after YAMSKIKH 1992). – 1 = loessy sand ; 2 = sandy loess; 3 = sandy slope loess; 4 = slope loess; 5 = semipedolite; 6 = gleyed silt; 7 = clay with boulders; 8 = debris; 9 = broken stone with sand; 1 = alluvial sand; 11 = alluvial boulder beds with sand; 12 = sandstone; 13 = steppe-type chernozem soil; 14 = alluvial meadow soil; 15 = grey forest soil; 16 = $CaCO_3$ accumulation ; 17 = loess doll; 18 = krotovina; 19 = charcoal; 20 = macrofauna; 21 = shells of molluscs; 22 = unconformity in the profile; 23 = traces of non-linear erosion; 24 = traces of linear erosion; 25 = pseudomorphs along ice veins; 26 = deformation due to solifluction; SL = Salba soil complex; KT = Kurtak soil complex; grain size distribution (mm): A = clay (< 0,001); I = silt (0,001–0,01); L = loess (0,01–0,05); H = sand (0,05–1,0)

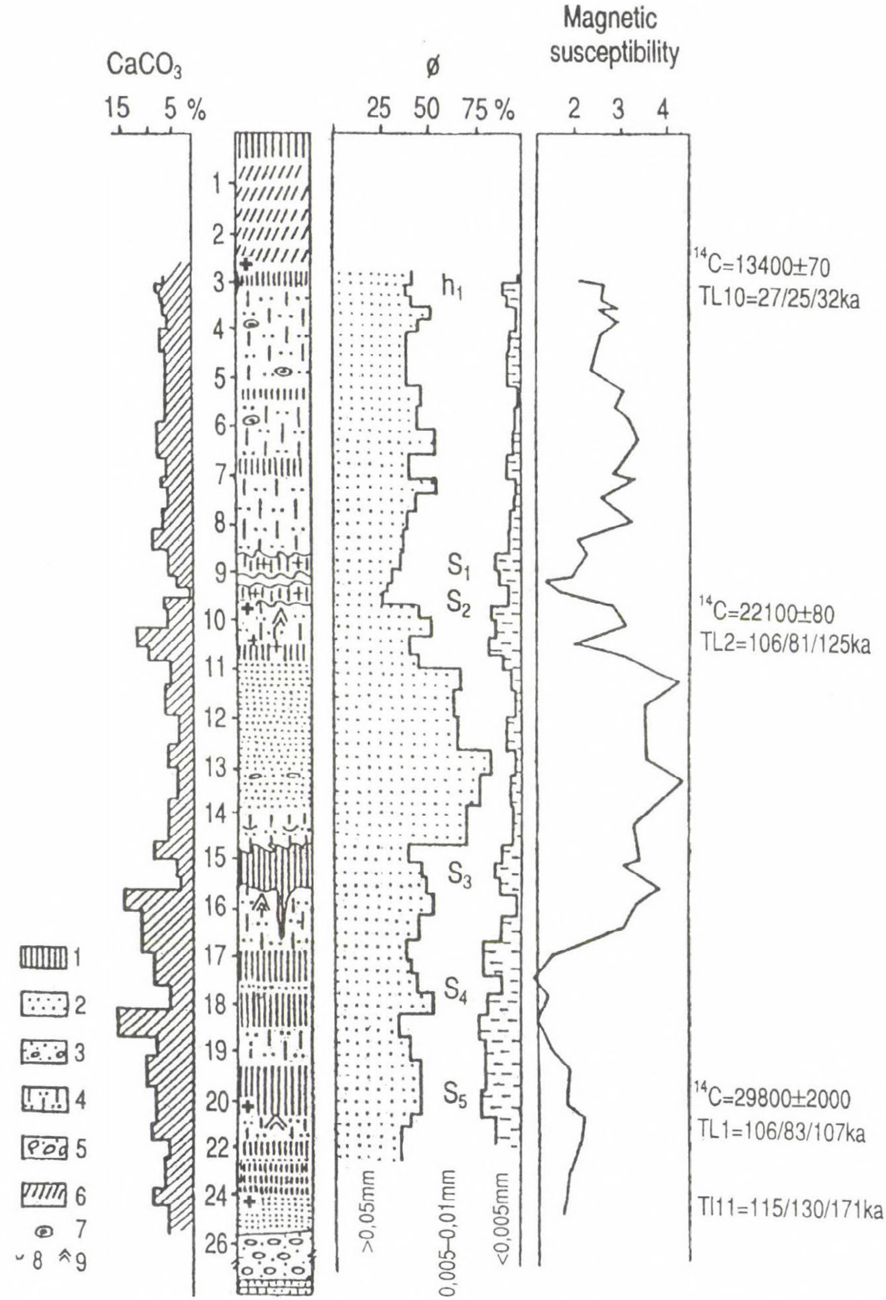


Fig. 6. Profile of the Lower Lagernaya (Tatyshev) terrace of the Yenisey at Krasnoyarsk (according to YAMSKIKH 1993). – 1 = soil; 2 = sand; 3 = gravelly sand; 4 = loam; 5 = gravel; 6 = slope loess; 7 = iron concretions; 8 = ferruginisation; 9 = Ca accumulation

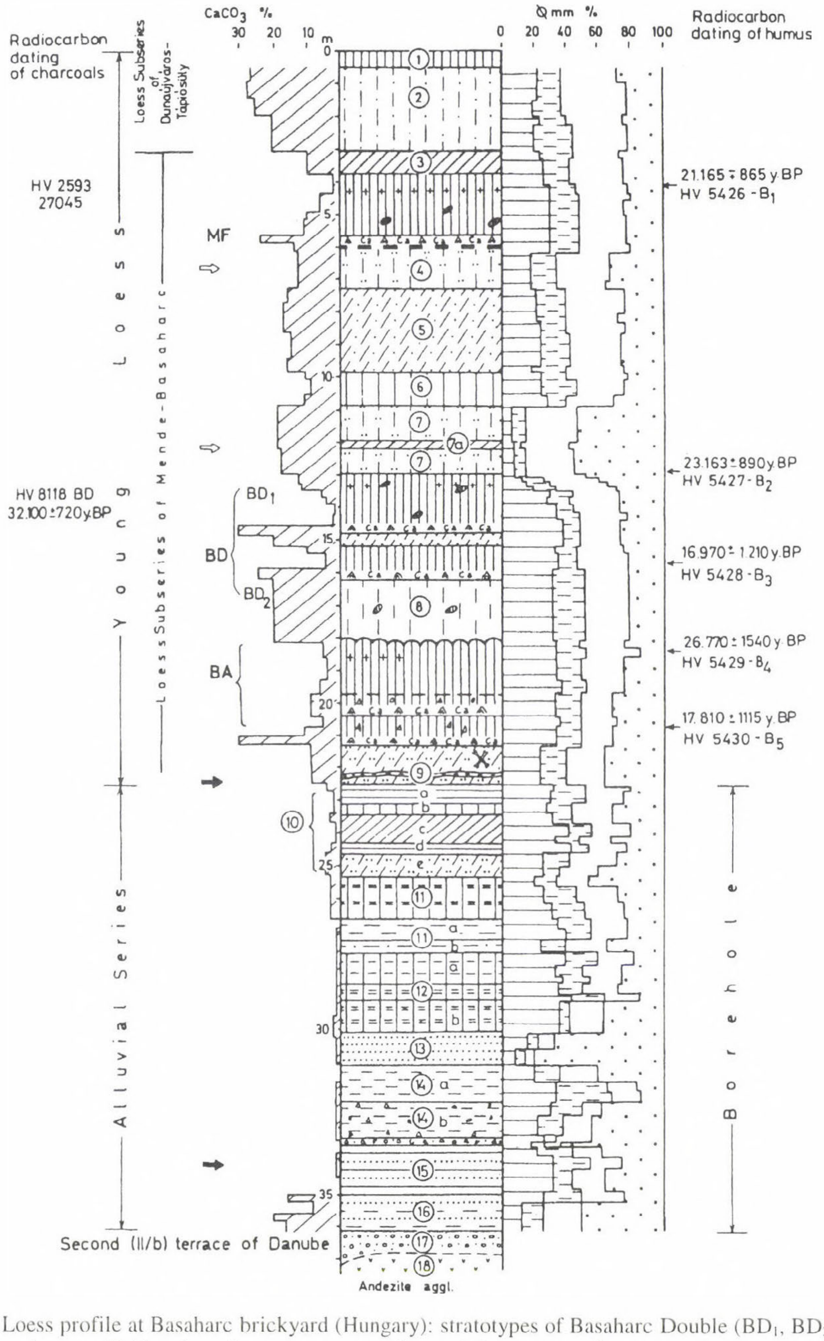


Fig. 7. Loess profile at Basaharc brickyard (Hungary): stratotypes of Basaharc Double (BD_1, BD_2) and Basaharc Lower (BA) paleosols. – + + + = charcoal fragments; ➔ = derasional (slope wash) unconformity; ⇨ = minor erosional unconformity

Young loess-paleosol-sand sequence of the lower Lagernaya (Tatyshev) terrace at Krasnoyarsk

The Paleozoic rock basement of the lower Lagernaya (Tatyshev) terrace is superimposed by a 24 m sequence of loess, sand and paleosols.

The basement of the rock is situated 9 m higher than the middle level of the Yenisey. In the studied exposure it is overlain by a 1.5 m thick terrace gravel and 1–1.5 m fluvial sand with thin interbeddings of sandy clay layers. From the lower part of this layer samples were taken for TL analyses *(Fig. 6, Table 2, Sib11).*

– 22–21 m sandy silt with two embryonic, hydromorphous, slightly calcareous flood plain soils of light brown colour and of 15–30 cm thickness each;

– 21–20.30 m crumby, calcareous, sandy loess, partly with a well developed C_{Ca} level of carbonate accumulation;

– 20.30–19.30 m meadow chernozem (S_5) of high sand content; sample Sib1 was taken from here;

– 19.30–18.60 m C_{Ca} horizon of carbonate accumulation with carbonate concretions and loessy sand;

– 18.60–16.90 m dark chernozem soil (S_4) of ochre-brown colour, strongly and medium calcareous;

– 16.90–16.00 m sandy loess, strongly calcareous;

– 16.00–15.50 m horizon of carbonate accumulation(C_{ca});

– 15.50–14.40 m chernozem soil (S_3) of high sand content and dark ochre-brown colour;

– 14.40–10.80 m loessy sand, sand;

– 10.80– 9.80 m calcareous sand, with a level of carbonate accumulation (C_{ca}) between 10.20–10.50 m (Sib2 sample from 9.80);

– 9.80– 8.40 m double meadow soil (S_1+S_2) of black colour (with a level of carbonate accumulation (C_{ca}) between 9.10–9.30 m);

– 8.40– 7.00 m sandy, with two intercalated tephras;

– 7.00– 6.80 m humous sandy loess (h_2);

– 6.80– 3.00 m sandy loess, loessy sand, between 5.20–5.35 m slightly humous sandy loess (h_{2sl});

– 3.00– 2.00 m two distinct layers of humous loessy sand (h_1) with a slightly calcareous accumulation, sample Sib10 was taken from here;

– 2.00– 0.00 m artificial infilling and soil developed on slope deposit covered by recent chernozem soil.

Radiocarbon data on *Fig. 5e* and *Fig. 6* (YAMSKIKH 1993) were based on ^{14}C datings of humus and an analogy (age of Ust'-Batoi horizon being 13.4 ka).

Experience has repeatedly shown that radiocarbon age of humus within loess-paleosol sequences is much less than that obtained by charcoal or mollusc shell analyses (see PÉCSI, M. and Hahn, Gy. 1987: loess-paleosol sequence at Basaharc, *Fig. 7*).

^{14}C analyses of charcoals usually do not exceed 30–35 ka and this is to be considered as minimum age. Absolute geological age might be underestimated twice or several times more.

Samples collected for TL and IRSL investigations from four representative horizons (SIB11, SIB1, SIB2, SIB10, see *Fig 6* and *Table 2*) were analysed by M. FRECHEN, Laboratory of Quaternary, University of Cologne.

Table 2. IRSL and TL dating of loess-paleosol sequences along the middle Yenisey Valley; age in ka (analysed by M. FRECHEN, Laboratory of Quaternary, University of Cologne)

Sample	IRSL/REGEN	IRSL/ADD	TL/REGEN	TL/ADD
SIB 92-1	87.9±8.7	106.9±14.5	88.7±12.2	107.2±11.5
SIB 92-2	76.1±7.7	106.4±19.9	81.7±8.7	125.6±13.0
SIB 92-3	62.2±7.3	99.0±25.6	71.2±9.2	102.2±10.1
SIB 92-4	20.9±2.4	23.1±2.7	23.3±2.9	20.9±4.6
SIB 92-5	79.8±9.5	121.9±16.9	15.7±9.8	112.6±11.6
SIB 92-6	19.0±2.3	21.9±4.4	28.4±5.2	30.3±5.5
SIB 92-7	76.4±8.7	104.5±15.0	86.2±10.3	123.8±12.0
SIB 92-8	55.5±5.6	65.4±8.4	60.4±7.2	62.2±6.1
SIB 92-9	75.5±8.6	96.7±11.5	90.3±10.1	110.5±13.7
SIB 92-10	22.5±2.4	27.0±5.4	25.2 ±3.1	32.2±4.2
SIB 92-11	96.4±11.6	115.6±19.3	130.7±16.0	171.1±16.7

Places of sampling:

1 – Tatyshev terrace, Krasnoyarsk
2 – Tatyshev terrace, Krasnoyarsk
3 – Cholnokov profile in the vicinity of Krasnoyarsk
4 – Sisim loess-paleosol profile along the middle reach of Yenisey
5 – Sisim loess-paleosol profile along the middle reach of Yenisey
6 – Primorskoe 1 loess-paleosol profile along the middle reach of Yenisey
7 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
8 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
9 – Primorskoe 2 loess-paleosol profile along the middle reach of Yenisey
10 – Tatyshev terrace, Krasnoyarsk
11 – Tatyshev terrace, Krasnoyarsk

Young loess-paleosol series on some old terraces of the Yenisey

A young loess-paleosol sequence overlying higher terraces of the Yenisey has proven to be very similar to the one on the Tatyshev terrace (see *Fig. 9*, Cholnokov loess exposure on terrace N°V). *Fig. 10* (Sisim) also comprises a young loess series being a retreating loess bluff above the dammed water level of the Krasnoyarsk reservoir in the vicinity of Primorskoe village. The sections mentioned are presumably part of the young loesses superimposing terrace N°V (in the vicinity of the paleontological and archeological site Primorskoe 1).

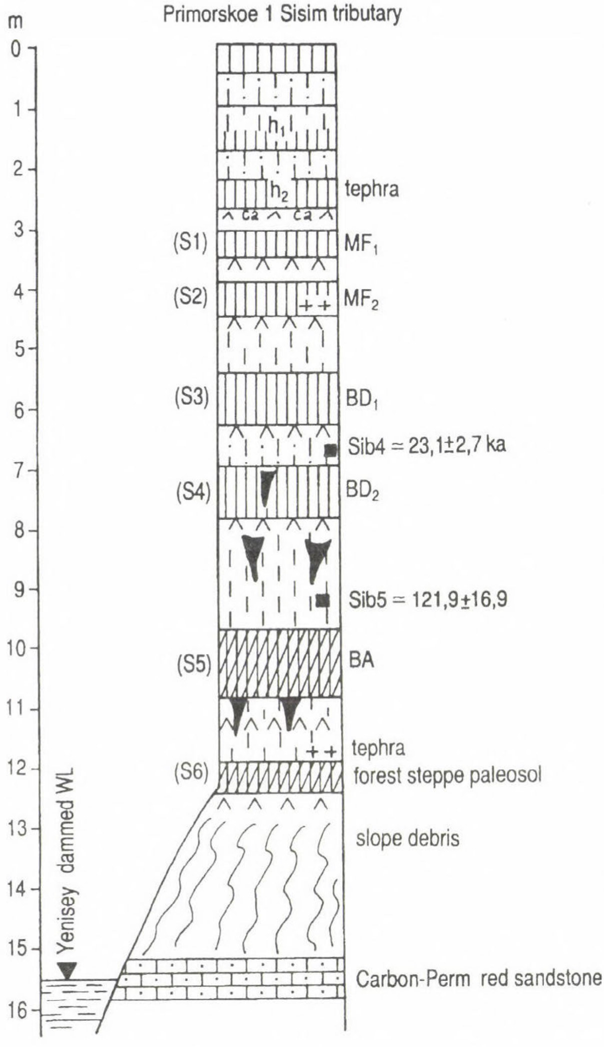


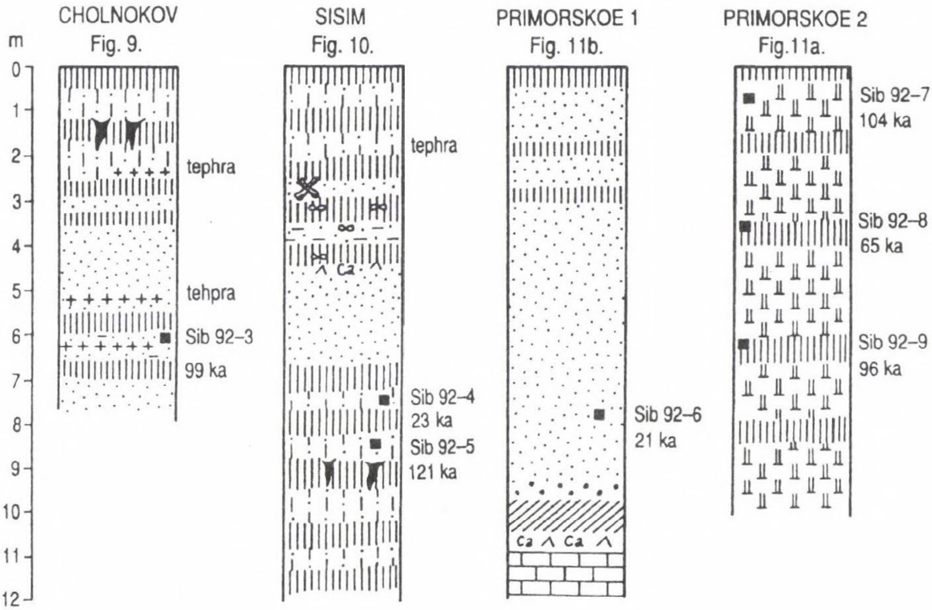
Fig. 8. Primorskoe 1 profile (Sisim – tributary of the Yenisey)

In *Figs 9* and *10* white tephra interbeddings are confined to five soil horizons. Sample SIB 92-3 was taken from a chocolate brown double chernozem; its age calculated by FRECHEN is TL/102 ka (IRSL/ADD calculation: 99.0±25.6 ka). According to its paleopedological type and stratigraphic position this double paleosol represents BD_1–BD_2 soil complex intercalated by a tephra layer of 10–15 cm thickness.[6]

Sample SIB 92-5 (Fig. 10) was dated 121.9±16.9 ka (IRSL/ADD). Stratigraphic position of this sample suggests a sandy loess with some carbonate accumulation between soils BD_2 and BA. Dating of sample SIB5 provides a good correlation but sample SIB 92-4 taken from between paleosols of BD_1 and BD_2 type and position shows an age of 23.1±2.7 ka. Perhaps this strong discrepancy is caused by carbonate accumulation or some other unknown effect.

In case of section Primorskoe 2 (Figs 11a and 11b) SIB-6=21.9±4.4 ka; SIB-7=104.5±15.0 ka; SIB-8=65.4±8.4 ka; SIB-9=96.7±11 ka. Reliability or weakness of results might depend on the fact that in profile 11a samples originate from old loesses while in profile 11b between the remnants of the old and eroded loess there is an interbedding of (polycyclic) fluvial sand and in the

[6] Denominations BD_1 and BD_2 (Basaharc Double) were used because position and genetic type of the Siberian paleosols show strong analogies with the third and fourth (double) steppe soil complex within a sequence overlying the flood-free terrace N°II of the Danube in Hungary (Pécsi and Hahn 1987, Pécsi 1996)



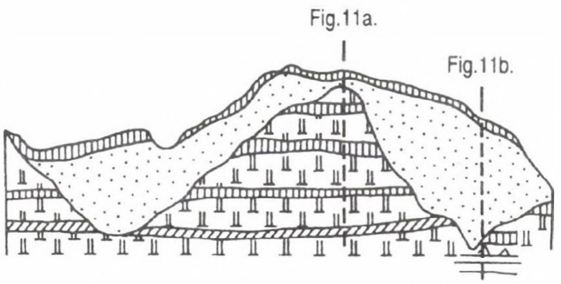


Fig. 9. Cholnokov loess-paleosol profile
Fig. 10. Sisim loess-paleosol profile
Fig. 11a and *b.* Primorskoe loess and sand bluff long the Yenisey at the Krasnoyarsk reservoir. Lithostratigraphy by PÉCSI, M., SCHWEITZER, F. and YAMSKIKH, A.F., TL/IRSL dating by FRECHEN, M.

uppermost part thin layers of slightly humous sand occur. Similar lithostratigraphic position is often very difficult to recognise because pillars of the old loess remnants in several places are covered by young but rapidly deposited sand. This young sand layer locally underwent weak loessification.

Conclusions

Terrace and loess-paleosol profiles presented in Discussion, and their lithostratigraphic evaluation, using the traditional methods of terrace chronology and loess-paleosol stratigraphy allow an approximate and so called relative chronostratigraphic subdivision involving bio- or lithostratigraphic schemes, time scales of climate fluctuations and calculating with the rate of landform evolution and deposition.

Absolute chronological methods are to complete this procedure. In our case two kinds of radiometric methods were available to assess the absolute age of certain key profiles.

Ages obtained through radiocarbon analyses considerably differ from those using TL or ISRL investigations. Radiocarbon datings of humus extraction provide young age for old soils. On the other hand ^{14}C ages represent a minimum age terminating at 25–30 ka. This method of dating is unable to determine older ages.

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RECONSTRUCTION OF SOME HOLOCENE GEOMORPHIC PROCESSES IN HUNGARY

GYÖRGY LOVÁSZ[1]

Introduction

Ca. 16 per cent of the 93,000 km^2 area of Hungary is covered by blown sand periodically accumulated since the end of the Pleistocene. Hills also occupy large areas (approx. 11 per cent). The valleys in the area of uplift since the Late Pannonian are cut mainly into Upper Pannonian sandy clay, and partly into Oligocene Schlier. The Upper Pannonian and Oligocene deposits are prone to sliding. On ca. 27 per cent of the area blown-sand accumulation as well as mass movements and sheet wash are the prevailing geomorphic processes.

The studies of landslides and blown sand features have a great tradition in Hungarian geomorphological research. The processes and the stages of development of the features are well known.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 015482.

Discussion

Landslides were most active during the Preboreal stage (10,200–9,000 yr B.P.) of the Holocene epoch (JUHÁSZ, Á. 1972., SZABÓ, J. 1996.). Three major blown sand accumulation periods are known. The first one is the last stadial of the Würm glacial (Late Gacial or Upper Pleniglacial). The movement took place around 20,000 B.P. The second stage was the Boreal (9,000–7,500 yr B.P.) and the third one occurred under the Ottoman rule, in the 16th and 17th centuries A.D. This last period is considered to be of human origin. The local population fled into the swamps and forests from the Turkish conquerors. As a consequence, sands started to move on the abandoned farmlands (BORSY, Z. 1977).

According to the investigations of Holocene vegetation and paleoclimate (ZÓLYOMI, B. 1952, JÁRAINÉ KOMLÓDI, M. 1966, 1969, KORDOS L. 1977, 1988, etc.), the periodicity of the landslides, migration of sand and sheet wash caused by precipitation can be traced well. These processes are largely dependent on climate.

The climate changes of the last 2000 years can be mainly studied and restored from historical records and dendrochronological investigations (RÁCZ, L. 1990, 1993, GRYNAEUS, A. 1997, LAMB, H. H. 1982, PFISTER, C. 1994, etc.). The last 160 years,

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i.e. the climate changes from 1841 until present can be best reconstructed from the data series of the Hungarian Meteorological Service. The geomorphological processes of the last 10,200 years can be investigated by the methods mentioned above.

On the basis of the investigations on vegetation history the Preboreal stage can be dated between 10,200 and 9,000 B.P. During this roughly 1000 years climate was cool and dry considering the distribution of pollen of the climate-indicator plants. In this stage – previously labelled 'spruce-birch' stage – the sheet wash had minor impact on the topography. It is also confirmed by the assumed 400 mm mean annual precipitation. Due to the low rainfall landslides did not occur frequently, although evaporation was also considerably low. Since climate was similar to present-day tundra conditions, the sand movement could not be intense on the wet surface resulted from the low evaporation rate.

The Boreal stage lasted longer, from 9,000 to 7,500 B.P. Climate was not uniform all through this 1500-year interval. Only the vegetation of the second half of the interval showed warm and dry character, when climatic conditions of intense sand motion were provided. Due to the lack of the precipitation water erosion did not play an important role with the exception of scattered heavy showers characteristic under steppe climate. Considering landslides the conditions must have been even less favourable for triggering them at that time than before.

The 2500-year interval of the Atlantic phase (7,500–5,000 B.P.) was the Holocene climatic optimum in the Carpathian Basin. It is indicated by the dominance of oak pollen. The precipitation radically increased in this period and climatic conditions were ideal for intensive sheet wash. There was a substantial deepening of valleys, and landslide processes gained in intensity. However, high annual precipitation did not allow large-scale sand movements.

During the 2500-year long Subboreal phase (5,000–2,500 B.P.) considerable climatic change occurred. At the beginning the climate substantially deteriorated, which is evidenced by the occurrence of spruce, fir and larch pollen. Ideal conditions for landslides prevailed at that time. Sheet wash and blown sand movements were not particularly intense.

According to the pollen analytical investigations, warming and increasing of precipitation happened ca. 3,000 years ago. At that time climatic conditions began to favour sheet wash compared to the former phase. The inclination to landslide increased in spring and autumn. The conditions of sand migration did not improve radically.

At the end of the Subboreal phase, ca. 2,750 years ago, another cool spell is indicated by pollen analysis. No data are available on the amount of precipitation, but it is likely that climate did not favour sand motion. The conditions and geomorphic processes at that time were similar to those of the Subboreal phase. The climatic deterioration was followed by rising temperatures at the beginning of the Subatlantic phase (2,500 B.P. to present). The Mediterranean climate probably supported wet winters. In this so-called Roman climatic optimum stage conditions for sand movements were favourable during the dry and warm summers. The humid summers and intense showers enhanced the processes of sheet wash.

The most characteristic climatic deterioration in the Holocene lasted for 300 years according to LAMB H. H. (1982), and for 560 years by PFISTER Ch. (1984). Both researchers agree that this so called 'Little Ice Age' terminated in the middle of the 19th century.

The climate of Hungary then bore resemblance to that of the glacial stages of the Pleistocene. As a result of the moderately wet and cold winters the inclination to landslides increased. It is likely that the freeze-thaw alternations accelerated infiltration and triggered landslides. Although the mentioned favourable climatic conditions existed during the cool and wet summers, landslide activity was reduced by increasing water loss by evaporation. Sheet wash was limited to short and wet summers. These climatic conditions were extremely unfavourable for sand movement.

The climatic oscillations of the past 160 years (from 1841 to present) can be best investigated with the help of meteorological data provided by the National Meteorological Service. The tendencies in the alterations of summer and winter half-year temperatures and precipitation were studied by trend analysis. Since the geomorphic processes highly differ in summer and winter, their separate investigation seems to be essential. Landslides are particularly common increasing principally in early and late winter, under wet and mild conditions. Under favourable climatic conditions the summer half-year was characterised by sheet wash and sand movement. Trend analysis, however, presents tendencies through ten-year moving averages and conceals possible extremities. As confirmed by present-day experiences, the intensity of geomorphic processes may substantially vary in years deviating from the trend.

During the past 160 years four distinct intervals can be identified, when climatic conditions in the summer and the winter half-years considerably differed from the previous or following intervals.

The 20 years between the mid-1850's and mid-1870's were characterised by warm and dry Mediterranean summers. The conditions for sand migration were favourable at that time. Sheet wash was uncommon during these dry summers. The winter half-years, however, were milder and drier. These conditions did not promote landslide generation.

In the 30-year interval from the mid-1870's to the beginning of the 1900's the summer half years were moderately cool and early in this period they were extremely, later moderately wet. The Atlantic summers were typical of the first half of this stage but hardly occurred in its second half. Since soils were wet for a long time mainly in the first half of the interval, no major sand migration can be assumed. These processes accelerated in the second half. In this phase the winter half-years were cooler and more humid than the average. It is likely that these conditions were suitable for landslides.

From the beginning of the 1900's until the mid 1940's summer half-years were mainly slightly cooler and more humid than the long-term average. In these Atlantic summers the intensity of sheet wash slightly increased. In contrast, climatic conditions for sand migration were not appropriate. During this interval of nearly 50 year mean winter temperatures were above the long-term average. Annual precipitation was below the average in the first part of this interval but rose above the mean values by the second half of the 1940's. During this relatively long interval winters changed from semicontinental to meditteranean. The climatic conditions became more suitable for landslides.

From the 1950's onward the summer half-years have cooled down considerably and annual precipitation has dropped. Summers have been rather cool and dry, reducing opportunities for sheet wash. Sand migration may intensify. Winter temperatures have shown a decreasing tendency but remain above the long-term average. Annual precipitation has notably reduced compared to the long term average, although short increases occurred occasionally. These climatic modifications indicate that the inclination to landslides has been diminishing for the last 50 years. Although, as it was already mentioned above, due to the smoothing character of the trend analysis, extremely humid or arid years could occur parallel to these tendencies, when the reduction or intensification of landslides might be typical.

Summary

The evaluation of the vegetation and climate history and meteorological data from the geomorphological aspect provided an excellent opportunity for the temporal analysis of the sand migration and sheet wash processes. It is likely that suitable conditions for sand migration were restricted to the second half of the Subboreal phase. The occurrence of landslides considerably increased during the Atlantic phase and in the so-called Little Ice Age.

The analysis of the maritime, continental and mediterranean oscillations of the last 160 years contributed to our understanding of the temporal alterations and provided valuable information for recent variations in the intensity of geomorphic processes.

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ENVIRONMENTAL CHANGES DURING THE LAST-, LATE- AND POST-GLACIAL IN HUNGARY

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Introduction

The palaeoenvironmental changes of the Quaternary can be traced in terrestrial deposits. During this period the climatic changes controlled the most important fluvial and aeolian processes such as surface modelling and sedimentation. The investigation of the loess profiles and sand dunes stratification provides data concerning palaeoenvironmental conditions in Hungary (*Fig. 1*). The Pleistocene loesses of the Carpathian Basin developed under cold steppe conditions, while the paleosoil horizons reflect a warmer and more humid environment. The climatic fluctuations affected the sandy material; the strong winds moved the sand during the dry phases and the vegetation gradually extended over the sandy surfaces during the wetter and milder periods. Below this vegetation cover steppe-like soil formed and the dunes had become fixed. The geomorphological effects of the climatic changes on the rivers were twofold:

– the fluctuations of the discharge of the rivers caused variation in the size of the meanders. Based on the different size and age of the abandoned river branches the Late-Glacial and Holocene development of the alluvial fans of the Great Hungarian Plain can be established and the palaeohydrology of the area reconstructed;

– the changing regimes of the rivers resulted in the formation of terraces and flood plain systems.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 022712.

Last-Glacial

As it is well-known loesses of the Carpathian Basin developed under cold steppe conditions, while paleosoil horizons formed in a warmer and more humid environment. A more detailed subdivision of the Pleistocene is possible using different methods, such as malacology, pollen and phytolith analysis, sedimentology etc., but for the interpretation of all these results it is necessary to know the chronological frame of

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Fig. 1. Location of the investigated loess profiles and sand dunes. – I = Basaharc; II = Paks; III = Mende; IV = Albertirsa; 1 = Kisrozvágy; 2 = Aranyosapáti; 3 = Székely; 4 = Nyírmihálydi; 5 = Vajdácska; 6 = Bodroghalom; 7 = Kenézlő; 8 = a site SSE of Debrecen; 9 = Dunavarsány

these events. The Last-Glacial represents a favoured domain in this respect, because the most direct or indirect "absolute" age determination range falls in this time interval. Loess and loess-like sediments cover more than 30 per cent of Hungary; so the investigation of loess sequences is a paramount question in Quaternary research. During the past two centuries numerous researchers investigated this sediment and created theories about the origin of the parent material, the age and circumstances of deposition.

The most important loess section is situated in the brickyard of Paks, where the first investigators (described in KRIVÁN 1955) tried to reconstruct the palaeoclimatic changes having occurred during the Pleistocene. KRIVÁN (1955) gave the first detailed and reliable description of Paks based on lithological, mineralogical, geochemical and biostratigraphical evidence, and carried out a palaeoclimatic reconstruction providing "absolute" ages using the MILANKOVITCH-BACSÁK astronomical curve. Subsequently PÉCSI (1966, 1979, 1987) worked out a classification of loess and loess derivates of Hungary and subdivided the Hungarian loess-paleosoil sequences into "Young Loess Series" and "Old Loess Series" based upon palaeoecological and lithological investigations and made a comparison of several loess sections. According to his nomenclature these two loess series are separated by the Mende Base (MB) paleosoil. This fossil soil played an important role in chronostratigraphy: based on western European experience – this is the first brown forest soil from the top – it was determined as the soil of the Last Interglacial (PÉCSI 1966, 1982, 1987).

In addition a synchronous volcanic intercalation was discovered first in the Paks profile by KRIVÁN (1959) and later they became known from other outcrops as well (JUVIGNÉ *et al.* 1991) in the loess between the Basaharc Lower (BA) and the MB fossil soils. This horizon, called Bag Tephra (BT), already suggested a much older age for the comprising loess than it had mentioned in the 1980's and 1990's. Recently the Bag Tephra is found at twelve sites in the Carpathian Basin and it is used as a stratigraphic marker horizon even if its age is not exactly determined yet (JUVIGNÉ *et al.* 1991, HORVÁTH *et al.* 1992, POUCLET *et al.* 1999). An other, much younger volcanic intercalation, the Paks Tephra (PT) was described from the northern wall of the Paks brickyard above the Mende Upper (MF) paleosoil (HORVÁTH 1999 in press). A corresponding tephra horizon from other outcrops is unknown yet.

In the light of new dating methods (thermoluminescence dating [ZÖLLER *et al.* 1987 in: PÉCSI–RICHTER 1996, WINTLE–PACKMAN 1988, LU 1991, 1992 in: PÉCSI–RICHTER 1996] and AAR determination [OCHES – MCCOY 1995]) and based on the above mentioned tephrochronology it is already clear, that the lower boundary of the Last-Glacial is shifted from the MB paleosoil upwards along the profile.

A recent combined luminescence investigation (FRECHEN *et al.* 1997) has confirmed the assumption that the Last Interglacial is represented by the MF_2 paleosoil in the Hungarian loess series (*Fig. 2*). Above and below the MF_2 paleosoil there is a significant sedimentation gap. The explanation for this hiatus could be the very low accumulation rate of loess during this time interval, early in the last glaciation. Very similar chronostratigraphy was established for loesses in Bohemia, Moravia and Slovakia (ZÖLLER–OCHES–McCOY 1994; OCHES–MCCOY 1995; FRECHEN *et al.* 1999). According to the above mentioned luminescence dating, the PK III paleosoil – which might correspond to the MF_2 paleosoil in Hungary – is situated within a time gap of ca 70 ka. It suggests a regional climatic impact for loess formation rather than a period of long erosion similar to some found at distant sites. This assumption correlates well with the oxygen isotope curves, where the characteristic cold peaks appear in the Upper Pleniglacial. A very high accumulation rate is present in the uppermost part of the investigated Hungarian loess profiles, as well as in the Czech loess sections (FRECHEN *et al.* 1999), which demonstrate a much colder and rigorous period in the palaeoclimate.

This investigation, which can deliver more precise results with the simultaneous use of the TL and IRSL methods (the IRSL signals are more accurate, while the TL ages are often overestimated, because the TL signals are less sensitive to bleaching and longer sunlight exposure is necessary), were carried out in three key stratotype loess profiles located at Paks, Mende and Basaharc, and further at a loess outcrop at Albertirsa; studies of the latter are published in the present paper for the first time (*Fig. 1*).

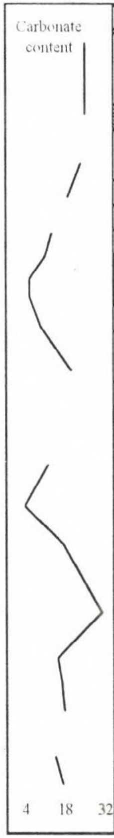
Description of the Albertirsa loess profile

This profile investigated lately – a 12 m high loess "wall" – is located in the Gödöllő Hills about 23 km southeast of the Mende section. There are two well developed buried soils, and an incipient humic horizon in the profile *(Fig. 3)*. This section belongs to the "Young Loess Series".

Hungary		Paks		Basaharc		Mende		Interpretation	
		TL (ka)	IRSL (ka)	TL (ka)	IRSL (ka)	TL (ka)	IRSL (ka)	Age 10^3 yr	^{18}O stage
		25,4±4,1 54,8±6,6	13,4±1,4 19,2±1,7					13–20	2
	h_1								
	h_2								
XXXX	PT			31,7±3,6	28,2±4,4	39,5±17,2 53,0±9,9	27,0±2,8 34,1±3,2	25–35	3
	MF_1								
						58,6±5,2	55,0±5,4	50–60	4
	MF_2								5
		101±10 168±17	120±12 170±24	153±14 208±21	121±14 156±46	206±21	149±14	> 130	6
	BD_1								
		177±16	144±15	184±16	131±28	149±14 231±20	138±14 162±15		
	BD_2								
		160±16 185±15	100±17 154±14	209±19 227±20	139±13 182±34	154±14 188±17	125±14 228±25		
	BA								
XXXX	BT	206±18 288±27	133±17 207±18	168±15	137±41	232±12 345±30	166±16 250±25		
	MB_1								
	MB_2								

Fig. 2. TL/IRSL ages of three Hungarian loess profiles above MB fossil soil (after FRECHEN *et al.* 1997)

The upper part of the section is 4 m thick and it mainly consists of sandy loess divided into two parts by the humic horizon. This youngest part of the section provides the most suitable matter for luminescence dating, so 8 samples (HCB 17–HCB 10) were collected from here. There is no substantial difference between the luminescence ages of the upper and the lower samples, so this thick loess sequence is assumed to have accumulated about 18,000–25,000 years ago, during the oxygen isotope substage 2b–2d. This indicates fairly high rates of deposition at that time, similarly to the other



Section		Sample	TL/additive age	TL/regeneration age	IRSL/additive age	IRSL/regeneration age	Munsell colour /dry/	Munsell colour /wet/
x	sandy loess	HCB 17	18348 ± 1958	21797 ± 2718	20685 ± 4927	20160 ± 3034	2.5Y 7/2	2.5Y 5/4
x		HCB 16	----------	----------	16794 ± 4763	----------		
x		HCB 15	22595 ± 2376	15679 ± 2039	16192 ± 14899	----------		
x	humus horizon	HCB 14	20122 ± 1760	18806 ± 2027	21716 ± 4788	17941 ± 2192		
x	loess	HCB 13	23381 ± 2047	25629 ± 12599	21898 ± 1990	21752 ± 2614	2.5Y 6/2	2.5Y 4/4
x		HCB 12	22758 ± 1936	21451 ± 3032	24357 ± 3068	20742 ± 2542	2.5Y 6/2	2.5Y 5/4
x	sandy loess	HCB 11	17161 ± 1557	20441 ± 2878	18193 ± 3294	17994 ± 2985		
x	sandy loess	HCB 10	20493 ± 1726	24859 ± 2536	22934 ± 2867	24183 ± 3051	2.5Y 6/2	2.5Y 4/4
	transitional horizon						10YR 6/3	10YR 4/4
	upper fossil soil						10YR 5/3	10YR 3/3
x		HCB 9	22666 ± 2893	22781 ± 2605	29109 ± 5157	24710 ± 3823	2.5Y 5/2	2.5Y 3/2
x	transitional horizon	HCB 8	37777 ± 3230	36640 ± 3558	37122 ± 4161	35432 ± 3865	2.5Y 6/2	2.5Y 4/4
x	loess	HCB 7	46622 ± 3948	----------	52998 ± 8686	----------	2.5Y 7/2	2.5Y 4/4
x	loess	HCB 6	47686 ± 4198	----------	50650 ± 18536	----------		
	transitional horizon						10YR 6/3	10YR 3/4
	lower fossil soil						10YR 4/2	10YR 3/4
x	transitional horizon	HCB 5	----------	----------	65931 ± 10675	64631 ± 10208	10YR 5/2	10YR 3/4
x	loess	HCB 4	124483 ± 10492	----------	120031 ± 18990	----------	2.5Y 7/2	2.5Y 5/4
	sand						10YR 5/4	2.5Y 4/4
x	loess	HCB 3	138706 ± 12375	----------	178844 ± 58398	----------	10YR 6/3	2.5Y 4/4
x	loess	HCB 2	180704 ± 15167	300723 ± 264000	126487 ± 18660	250508 ± 264646	10YR 6/3	2.5Y 4/4
	sandy loess							
	sand						10YR 6/3	2.5Y 4/4
x	sandy loess	HCB 1	88032 ± 7412	----------	92809 ± 27368	----------	10 YR 6/3	2.5Y 5/4

Fig. 3. TL/IRSL ages from Albertirsa loess profile

Hungarian profiles of the Upper Pleniglacial, but one difference is apparent: whereas the age of the loess in the Albertirsa section is 18,000–25,000 years, in other profiles (e.g. Mende) loess layers younger than 20,000 years are only found and they are absent between 20–25 ka. Accordingly loess accumulation was considerable during the whole Upper Pleniglacial, but – due to the uplift and subsidence of the different areas – part of the material sedimented was later eroded at some places.

The luminescence age of the humus horizon was found between 20 and 22 ka (after the additive ages), so it might be the h_2 layer. It was probably a warmer and wetter spell, which allowed the expansion of vegetation and soil formation, but the time period was not long enough to form a well-developed soil horizon.

The next 3 m thick portion of the section contains the first fossil soil. This 1.2 m thick upper paleosoil consists of brownish soil material and in its lower part crotovinas of pale colour are found. The colour of the soil gradually fades, its humus content decreases, whereas its carbonate content increases with depth. Below this first soil two thicker loess horizons, and a thin "transitional" layer towards the lower fossil soil are found. Samples were collected from each strata (HCB 9–HCB 6) except from that underlying "transitional" layer. This part of the profile comprises a longer period (about 25,000 years) than the former one. It could not be regarded as a complete sequence, because we must take into account an erosional gap might be assumed. The luminescence ages measured from the soil are rarely accurate, generally there is an underestimation. Therefore these ages had to be ignored during interpretation, and instead the ages of layers below and above the paleosoil were taken into consideration. So the first paleosoil can be identified with "Mende Upper 1" (MF_1) soil in other Hungarian profiles. During this interstadial period (Middle Pleniglacial) the climatic conditions promoted the expansion of cold forest steppe vegetation (pine [*Pinus silvestris*] and larch [*Larix decidua*] remnants were determined from the MF_1 fossil soil at Mende.) The luminescence ages of the two 1.5 m thick loess layers below the MF_1 soil were very similar (about 50 ka). The age of this loess sequence indicates probably a complete layer with high accumulation rates.

The thickness of the lowest part of the profile is more than 4.5 m. It consists of the lower paleosoil, and below that 6 loess and sand layers can be found. The soil contains a darker layer and a lower, lighter horizon. The upper part of the soil is richer in humus, and the lower one has higher carbonate content. The whole soil formation is densely intermingled by crotovinas, so this material is partially reworked. Luminescence samples (HCB 5–HCB 1) were collected from the loess underlying the lower fossil soil (HCB 5 was taken from the and, it is (similar to the upper paleosoil) ignored). The luminescence age of the loess layer below the soil was determined as more than 120 ka, therefore it can be ascertained that the lower soil corresponds to the Mende Upper 2 (MF_2) fossil soil and it was formed during the last interglacial (oxygen isotope substage 5e) or in an interstadial interval (5a/5c).

To sum up: the Albertirsa section contains the most complete loess sequence from the Upper Pleniglacial, a most developed paleosoil complex Mende Upper (MF_1 and MF_2) studied in Hungary and the h_2 humic horizon generally encountered in the Hungarian loess sections.

Late-Glacial

During the Late Glacial period the most important (fluvial and aeolian) geromorphic processes were governed by the climatic changes. Based on the investigations of the blown sand territory (dunes and sandy materials) and the fluvial terraces, alluvial fans of the region can present ideas for the description of past environmental conditions of Hungary.

The *blown-sand regions* of the Great Hungarian Plain originated from the material of alluvial fans built by the rivers running from the Carpathian Mountains into the basin. These watercourses deposited 10–400 m thick series of sediments (mainly sands and fine grained deposits) on top of the Pliocene strata. Climatic fluctuations during the Pleistocene already affected this sandy material; during dry periods the strong winds removed the sands from the surfaces with scant vegetation. During the wetter and milder (interstadial) periods the vegetation gradually closed on the sandy surfaces. Under this vegetation cover steppe-like soil formed and the dunes became fixed.

For dating aeolian sediments two different methods are used in Hungary. During the 1980's an indirect method was developed by BORSY *et al.* (1982). blown-sand Eight exposures contain about ten intercalated fossil soils in the north-eastern part of the country *(Fig. 4)*. Organic material of these fossil soils was analysed by the radiocarbon method. This dating of the deflation period is indirect, because the data relate to the soil formation and not to the aeolian phases. BORSY (1985) and LÓKI *et al.* (1995) succeeded in determining the age of blown-sand periods in the sand dunes and established a pattern for sand movement.

Under cold and arid climates prevailing during the Upper Pleniglacial the surface of the alluvial fans was attacked by strong winds, and this was the main deflation phase of the Hungarian blown-sand territories. The well developed fossil soils in the dunes indicate that during the Bölling and Alleröd phases the climate turned somewhat milder and more humid than before. A steppe-like vegetation cover evolved gradually on the surface of the previously moving sand dunes. Scrubs and spots of pine-birch forests also appeared (BORSY *et al.* 1985). Close to the eastern bank of the Danube – in the central part of the Carpathian Basin – poplar and willow groves occurred as well. During the warmest period of the Alleröd interstadial the mean temperature for July was about 4°C lower than today on the Great Hungarian Plain. The cool spells (Older and Younger Dryas) were characterised by intensive sand movements. The low temperatures (7–8°C below the present-day ones) and the limited precipitation resulted in the shrinkage of vegetation cover and the reoccurrence of sand movement and sand accumulation on the previous dune surfaces fixed by soils *(Fig. 4)*. This investigation on the aeolian sand areas in north-eastern Hungary does not provide any proof of sand movement during the Post Glacial (Holocene).

The second and new method of dating aeolian periods is direct: dating the blown sand by various luminescence methods. This allows a reliable determination of the time when climatic oscillations related environmental changes occurred by tracing them in the sand-paleosoil sections, because the absolute time of the aeolian events can

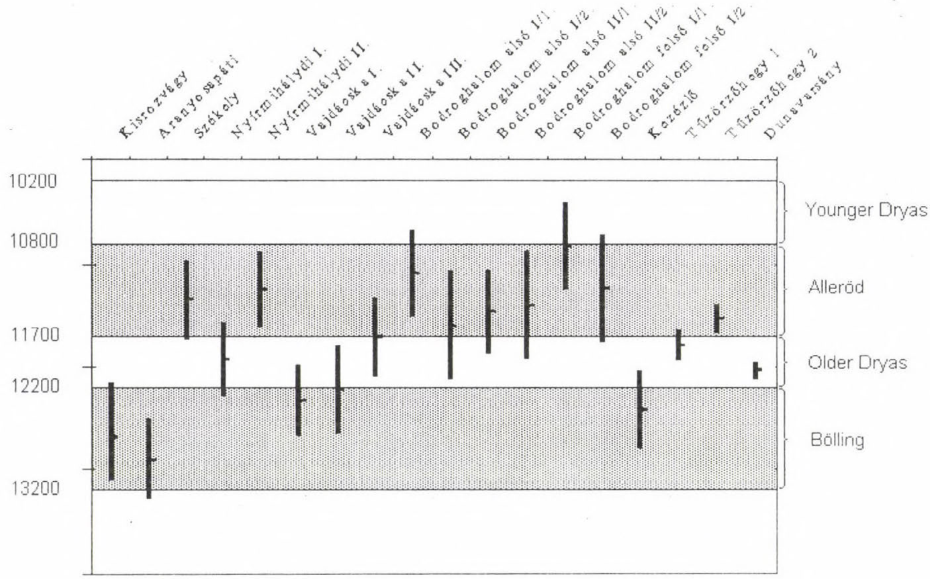


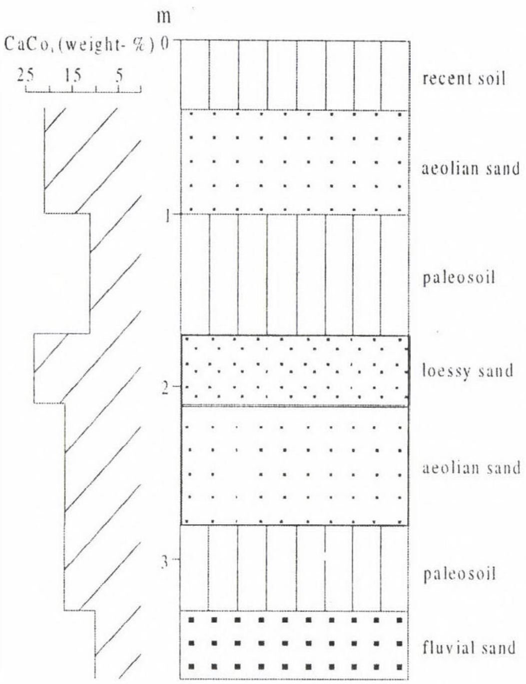
Fig. 4. Radiocarbon (C^{14}) data from Hungarian sand dunes

be determined directly. This method used for first time for aeolian sand in Hungary, and the results are demonstrated by the example of sand profile at Dunavarsány.

Description of the Dunavarsány sand profile

The abandoned sand quarry near Dunavarsány (at the far north-western part of the Danube–Tisza Interfluve) provides a good opportunity to investigate the surface processes of the Danube alluvial fan. The section lies 600 meters from the actual Danube bed and contains two paleosoils and several sand layers. The assumption that the humic layers represent humid periods of the Late-Glacial and/or the early Holocene, were confirmed by the results of luminescence and radiocarbon datings (*Fig. 5*). The material of the sand horizons originates from the alluvial deposits of the Danube: the grains were blown out by the north-western winds during drier phases. According to the electron-microscopic examination and the grain-size investigation the short distance of transportation permitted only slight abradation typical and classification of the grains.

The coarse sand formation constituting the base of the section contains thin grit layers too and it has the coarsest grain size at the profile (94% of the material falls above 0.1 mm, and 44% above 0.2 mm particle size). Its fluvial origin is verified by electron-microscopic examination which demonstrated that the surface of the grains have sharp fractures and edges, the abradation of the aeolian grains is missing completely (*Fig. 6a*).



Sample number	IRSL age (ka)	TL age (ka)	Radiocarbon age (ka)
HDE 07	6.0±0.5		
HDE 06	7.7±0.6	5.7±1.0	
HDE 11	8.1±0.5	7.8±1.0	
HDE 10	9.2±1.7	6.8±2.2	
HDE 9	12.0±1.9	9.6±1.4	
Deb-6440			12.04±0.06

Fig. 5. Sand dune profile with radiocarbon and TL/IRSL ages at Dunavarsány

Above the fluvial sand horizon lies a brownish, well developed 0.4 m thick paleosoil. Its humus content is 0.47%, while its carbonate content exceeds 16%. The charcoals found in this horizon – according to xylothomic analysis – are the remnants of poplar (*Populus sp.*) and willow (*Salix sp.*), while there is no evidence of the presence of fir species. These results point to a definitely mild, wet climate, which is also suggested by the thickness and the high humus content of the soil. The radiocarbon age of the charcoal is 12,040±60 years (Deb-6440).

The soil horizon is overlain by a dark-coloured 0.7 m thick sand layer, which becomes lighter moving downwards due to the presence of carbonate concretions. The average carbonate content varies between 15–17%, but this value is higher in the lower part of the layer. The grain size is ordinary, since 47% of the material falls in the 0.14–0.50 mm range. The luminescence age (HDE 9) of the layer is 12.0±1.9 (IRSL); 9.6±1.4 ka (TL). This result is in accordance with the radiocarbon age of the underlying paleosoil.

Above the sand horizon lies a 0.5 m thick loessy sand, which is the finest horizon of the site, because 21% of its material falls in the size range below 0.05 mm and 12% below 0.02 mm. The light colour of the layer is due to its high (25%) carbonate content. According to the electron-microscopic examination the marks of the fluvial action dominate on the surface of the grains, but numerous grains display traces of aeolian transportation as well. The fact, that no typical aeolian grains could be found, argues for the short duration of aeolian transportation. The luminescence ages (HDE 10) of the layer are 9.2±1.76 (IRSL) and 6.8±2.2 ka (TL).

The thickest buried soil of the section lies above the loessy sand. Its thickness (0.8 m) and dark colour – apparently originating from the high humus content (0.54%) – relates to a long and ideal climatic period favouring soil formation. The carbonate content shows significant horizontal variance, since its value ranges between 0.83 and 11.29%. This indicates heavy leaching, which – depending on the relief – washed the carbonate content into the underlying horizon to various degrees. The luminescence age of the layer measured on two samples (HDE 11 and HDE 6) is 8.1±0.5 (IRSL), 7.8±1.0 (TL) and 7.7±0.6 (IRSL), 5.7±1.0 ka (TL).

The uppermost sand layer of the section is situated between the recent soil and the upper buried soil. 78% of the material falls in the 0.1–0.5 mm range and the carbonate content exceeds 20%. Electron-microscopic examination revealed strongly abraded grains in this horizon (*Fig. 6b*). The luminescence age (HDE 7) of the layer is 6.0±0.5 ka (IRSL).

Interpretation of the results

According to the investigations described above, we can outline the process of dune formation, and work out a theory for the climatic change and geomorphologic processes having taken place in this region. The underlying fluvial sand of the dunes are the deposits of the Danube River, the channel of which was shifting gradually to the west during the second part of the Pleistocene. The first overlying paleosoil layer (according to radiocarbon dating) developed during the Bölling interstadial of the Late-Glacial. The mild, wet climate on the riverbank enabled the expansion of deciduous vegetation. Under the deteriorating climate during the Older Dryas the wind covered the soil with thick sand. On the top of the dune (which was exposed to the stormy winds, and therefore was not suitable for the settlement of any vegetation) no humic layer was formed during the Alleröd interglacial or it was completely eroded later. The luminescence age of the loessy horizon points to the fact that no continuous vegetation cover was developed on the top of the dune even during the early Holocene and the movement of sand was practically continuous. Hence during the Late-Glacial the airborne falling dust mixed with the sand and it continued moving until the Preboreal phase of the Holocene.

The uppermost paleosoil was being formed under the relatively mild climate during the first part of Atlantic phase in the Holocene. Beside the results of dating this is suggested by the high humus content and the considerable thickness of the horizon The uppermost sand layer was transported to its present location during the dry second part of the Atlantic phase (6.0±0.5 ka). This is the first well founded date of Holocene sand movement in Hungary. Moreover, this result is in accordance with other recent investigations, which proved dry conditions by the low level of the Lake Balaton, the minimum of the humidity curve (KORDOS 1977), and the occurrence of alkali soils.

The geomorphological effects of the climatic changes on the rivers were varied:

– the discharge fluctuations of the rivers caused variation in the size of the river meanders, and based on the different size and age of the abandoned river branches the Late-Glacial and Holocene development of the alluvial fans of the Great Hungarian Plain and the palaeohydrology of the area can be outlined;

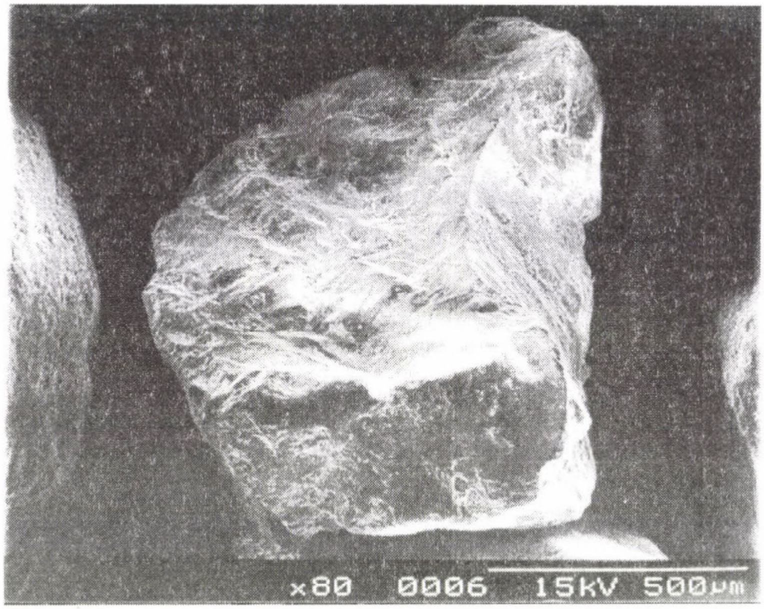
– the climate controlled changing mechanisms of the rivers resulted in the formation of terraces and flood plain systems.

A wide range of abandoned alluvial meanders has already been discovered on the Great Hungarian Plain. The abandonment of the channels has been dated using radiometric (^{14}C) dating methods, pollen analysis and geomorphological considerations. During the past decades a number of researchers conducted measurements and calculations concerning the statistical relationship between the discharge rates of meandering rivers and the size of the meanders. One of the authors has drawn relationship between the meander parameters of active meanders and the present discharge properties of various rivers on the Great Hungarian Plain (GÁBRIS 1985). Using the above mentioned river discharges the meander properties, and equations of the palaeodischarge could be calculated from the size of the meanders left by the ancient rivers (GÁBRIS 1987, 1995/a). The results provide quantitative evidence to the fluctuation of discharges of the palaeochannels during the last thirteen thousand years of the Late-Glacial and the Holocene, and enable an outline of the general palaeohydrologic pattern of the Great Plain (GÁBRIS 1995/b, 1998). On the basis of a relationship (NOVÁKI 1991) between runoff and two climatic elements (annual mean temperature and annual precipitation) the author attempted to determine annual average precipitation from Holocene discharges estimated by morphometric methods for some periods. These data have geomorphological consequences too, and these consequences will be discussed in the next section.

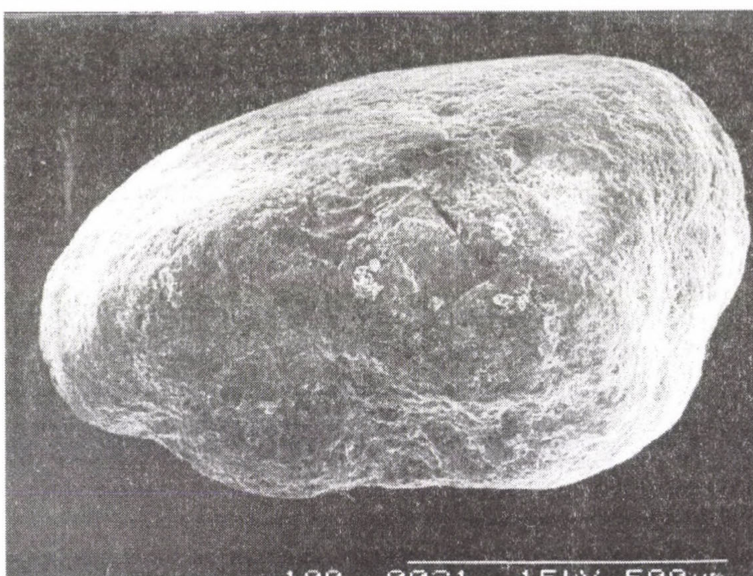
The youngest Pleistocene terrace (No. II/a) of the Hungarian rivers was previously dated to the Holocene because of the absence of a loess cover on its surface (PÉCSI 1959). The exceptional height above the present-day flood level was attributed to the sand dunes on the terrace. This sand accumulation was dated to the earlier Boreal. However BORSY *et al.* (1982, 1985) and our recent investigations proved that the main aeolian deposition phases (when sand dunes were formed on the flood-free surfaces of the II/a terrace) are attributed to the Pleistocene, especially to the Pleniglacial and Late-Glacial (Early and Younger Dryas). On the basis of the pollenanalytical results (JÁRAINÉ KOMLÓDI 1966) the Late-Glacial period is characterised by great variations in temperature and the water-level changes of Lake Balaton testifies on substantial variation of precipitation as well (NAGYNÉ BODOR 1988).

Post-Glacial

At the beginning of the Preboreal the rise of precipitation and temperature (the mean annual temperature was about only 1–2°C lower than the present-day one) pro



A



B

Fig. 6. SEM pictures – two type of sand grains from the Dunavarsány profile. – A = Quartz grain with sharp fractures and edges from the bottom of the profile – there is no evidence of aeolian transport; B = Well abraded aeolian quartz grain from the top of the profile

duced an increased water flow and a reduced sediment load; consequently a downcutting fluvial phase appeared for a short time.

Until recently the Boreal was considered to have been the driest phase during of the Holocene (JÁRAINÉ KOMLÓDI 1966) and the most significant sand movement during the Holocene was dated to this period too. But BORSY *et al.* (1982, 1985) could not prove sand movement during the Boreal in the middle and eastern Hungarian aeolian sand regions. Only a slight modification of the older sand dunes is traceable, but even this cannot be conclusively separated from the wind erosion effects due to the human intervention in the 17–18th centuries A.D. The relatively high water level of Lake Balaton also points to a wet climate. The palynological examinations, the "vole-thermometer" (KORDOS 1977) and the new "malaco-thermometer" methods (SZÖŐR et al 1991) all point to a climate warmer than today. According to the most recent data it is problematic to evaluate the environmental conditions of the Boreal period. The fluvial activity was changing and the accumulation reappeared during the second half of the Preboreal and the first half of the Boreal.

The next channel deepening phase can be placed at the Boreal–Atlantic transition (the erosion of the early Holocene terrace No. I), when the climate turned wetter. The high water level of Lake Balaton and the "humidity" curve (KORDOS 1977) both suggest that a wet period of the Hungarian Holocene occurred during the transition between the Boreal and Atlantic phases. However, the palynological examinations in eastern Hungary show considerable differences in the climate of this period in the Carpathian Basin: the eastern part of the Great Hungarian Plain has proven to be much more continental. The results of this wet phase were more frequent floods and extreme water discharges.

The ideas concerning the palaeoenvironment of the Atlantic phase have to be changed thoroughly in the light of recent investigations. The low water level of Balaton (CSERNY *et al.* 1991), the lowest point on the humidity curve (KORDOS 1977), and the results from pollen analysis (CSONGOR *et al.* 1982) have all proven that at least the second half of the Atlantic phase was much drier then hitherto thought of. The palaeohydrologic reconstruction by radiocarbon dating shows that the young Atlantic (6,000 B.P.) channels of the Tisza River (BORSY Z.–FÉLEGYHÁZI E. 1983) had a lower water discharge than today (GÁBRIS 1985, 1998). For the first time in Hungary recent investigation in the Dunavarsány sand quarry have demonstrated a deflation period: the uppermost aeolian sand layer of this sand dune was transported here during the dry second half of the Atlantic phase (luminenescence age is 6.0±0.5 ka [IRSL]). The stratigraphic evaluation of archaeological excavations (BÁCSKAI 1991) also indicates dry conditions: the buried alkali soil, which formed above the middle Neolithic cultural layer, points to an excessively continental climate, with a negative water balance. There are similar alkali soils in some Middle Copper age (approx. 4,500 B.P.) settlements, too. The continental character of the climate increased, and so did the frequency of flood, and the water yields became extreme. Thus the rivers were silting up during the second half of the Atlantic and the first part of the Subboreal periods as a result of the dry climate and low discharges.

The next erosion took place during the second, wet part of the Subboreal. KORDOS' (1977) "humidity" curve shows a highly humid period around 3000 B.P. and NAGYNÉ BODOR's (in CSERNY *et al.* 1991) pollen investigations of the eastern Balaton Basin also place the highest water level during the second half of the Subboreal. These data indicate that the latest phase of the Holocene downcutting was quite possibly not during the early Subboreal (SOMOGYI 1962), but during the second half of it. Due to the high rainfall and the low temperatures the rivers with abundant yield incised: an eroded scarp was formed separating the higher floodplain level from the lower floodplain. The Danube (and all the other Hungarian rivers) currently demonstrate a slight downcutting tendency.

Acknowledgement – Our investigation during the year 1999 was supported by the FKFP 0180/99 project of the Office for Higher Education. We thank Dr. Manfred FRECHEN for the facilities, the assistance and control of measuring the IRSL and TL ages of samples of the Albertirsa and Dunavarsány profiles at the Cheltenham Geochronology Laboratories. We are grateful to Dr. Károly BABOS (Eötvös University of Budapest, Dept. of Plant Anatomy) for the xylothomic examinations of the charcoals originating from the paleosoils of the Dunavarsány and Mende profiles.

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SOIL EROSION ASSESSMENT AND MODELLING

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Abstract

Soil erosion is a very important land degradation process in the hilly areas of Hungary. Soil erosion research dates back to the early 1950s when the whole country was mapped on the scale of 1:75 000 and the first attempts were made to assess soil loss by the USLE.

The paper presents three different methods of assessment and modelling soil erosion. The USLE was applied in a tributary catchment of Lake Balaton, the EPIC model for the same catchment and another application of the USLE occurred in the catchment of Lake Velence using a different algorithm. The aim of this study is to show three methods of soil erosion assessment and to compare their applicability.

Introduction

Soil erosion is probably the most important land degradation process on hillslopes. For many experts land degradation and soil erosion mean exactly the same thing and conservation is nothing else but measures to be taken against soil erosion. In reality land degradation includes several processes as it will be explained below. According to the definition of UNEP (1992), land degradation is the "... reduction of resource potential by one or a combination of processes acting on the land". JOHNSON and LEWIS (1995) define land degradation as "... the substantial decrease in either or both of an area's biological productivity or usefulness due to human interference". Though land degradation processes are induced both by natural and human factors the role of the human impact is definitely more important. This statement applies especially to those countries of the world where agriculture still has a considerable contribution to the national economy like in Hungary.

Land degradation processes include soil erosion by water and wind, chemical degradation (acidification, salinization/alkalization processes, leaching) as well as physical degradation of soils (soil compaction, crusting, structural damage, degradation due to the extreme soil moisture regime) and biological degradation. *Table 1* presents world-wide data on the extent of degraded land due to erosion.

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Table 1. Global extent of soil degradation due to erosion, by region
Source: OLDEMAN, HAKKELING, and SOMBROEK 1991.

Region	Area eroded by water erosion: Light	Moder-ate	Strong and extreme	Total	Area eroded by wind erosion: Light	Moder-ate	Strong and extreme	Total	Total area eroded	Total area seriously eroded	Total seriously eroded as a percent of total land used
	(million hectares)										
Africa	58	67	102	227	88	89	9	186	413	267	16
Asia	124	242	73	441	132	75	15	222	663	405	15
South America	46	65	12	123	26	16		42	165	93	6
Central America	1	22	23	46	246	4	1	5	51	50	25
North America	14	46		60	3	31	1	35	95	78	7
Europe	21	81	12	114	3	38	1	42	156	132	17
Oceania	79	3	222	304	16		27	46	99	3	3
World	343	526	223	1094	269	254	26	548	1642	1029	12

The objective of this paper is to give a short overview of soil erosion studies in Hungary followed by the presentation of three different methods of assessment using different models to test their applicability.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 024165, T 032274.

Soil erosion in Hungary

Soil erosion mapping started as early as the 1950's when the awareness of the damage due to erosion grew. After the first attempt by J. MATTYASOVSZKY (1953), who surveyed the western part of the country T. DUCK–P. STEFANOVITS constructed a 1:200,000 scale map of all the mountain and hill regions of Hungary (DUCK, T. 1960). More recent soil erosion surveys in Hungary are generally based on a conventional estimation of soil profile truncation: how deep soil layer is missing compared to an intact profile of the same type of soil in the region (STEFANOVITS, P. 1964).

Table 2. summarises the extent of soil erosion in Hungary. About two thirds of the total area of Hungary is under cultivation and for that reason agrogeomorphic processes including soil erosion are of great importance as regards the present-day surface evolution.

Table 2. Soil erosion in Hungary (after STEFANOVITS, P.–VÁRALLYAY, GY. 1992)

	Thousand hectares	% of total area	% of agricultural land	% of eroded land
Area of the country	9303	100	1	
Area of agricultural land	6484	69,7	00	
Arable land	4713	50,7	73	
Total eroded land	2291	24,7	35,3	100
strongly	554	6,0	8,5	24,1
moderately	885	9,5	13,6	38,5
weakly	852	9,2	13,2	37,4

The Universal Soil Loss Equation (USLE) was first used for soil conservation planning in the 1960's and 70's. In the Planning Office for Water Management (VIZITERV) a system was elaborated based on the USLE to provide quantitative data for planners (ERŐDI, B. *et al.* 1965, 1974) and this was accepted as the standard approach to soil conservation.

Soil erosion estimations from the equation gradually became part of erosion mapping (MÁTHÉ, F. 1974) and the prediction of erosion hazard was founded on meteorological and soil parameters. Recently, process monitoring on catchment scale (PINCZÉS, Z.–KERÉNYI, A.–MARTON-ERDŐS, K. 1978; GÓCZÁN, L.–KERTÉSZ, Á. 1988), testing of soil conservation technologies in agriculture (BIRKÁS, M.–SZABÓ, L. 1992) and laboratory experiments with raindrop impact (KERÉNYI, A. 1991) also led to important new results.

Soil erosion assessment in Lake Balaton Catchment

Assessment of soil loss by USLE

Lake Balaton Catchment lies in western central Hungary. The lake with a total area of 577 km^2 is exposed to various kinds of environmental impacts including agricultural activity in the catchment. The influx of sediment and solutes into the lake deriving mainly from *non-point pollution sources* plays an important part from the aspects of the *eutrophication* and pollution of the lake.

One of the tributary catchments on the northern shore of the lake was selected for closer study. A detailed analysis of the test area, the Örvényesi-Séd catchment (24 km^2), forms the basis of extrapolation for the northern catchment *(Fig. 1)*. The elevation of the highest point in the area is 416 m a.s.l., the outlet lies at 104 m a.s.l. (the relative relief is 312 m). The main stream, the Örvényesi-Séd is 8.1 km long. Cambisols developed on loess, otherwise rendzinas and vertisols are the main soil types of the area.

For the assessment of soil loss within the test area the Universal Soil Loss Equation (USLE) was applied. As it is well known, the USLE expresses the rate of soil erosion in the form of soil loss (t/ha), i.e.

$A = RKLSCP$

where A is the predicted soil loss (t/ha),

R is the rainfall and runoff factor,

K is the soil erodibility factor,

LS is the factor of slope length and steepness,

C is the cover and management factor and

P is the support practice factor.

ARC-INFO was used for data management and manipulation. Runoff directions and slope angles were calculated from the DEM by the application of a triangular network, thus enabling the delineation of the small territorial units of soil erosion assessment, the so called *erotops*. An *erotop* is defined by G. RICHTER as a unit with

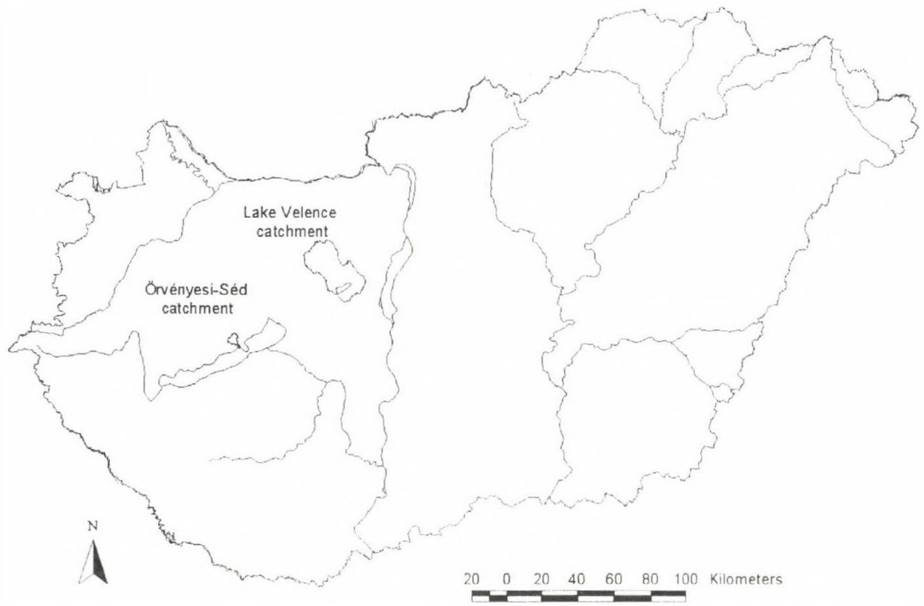

Fig. 1. Geographical setting of the catchments of Örvényesi-Séd and of Lake Velence in Hungary

approximatively the same runoff direction and without water collecting linear elements. They are bordered by the lines of diffluent or confluent runoff direction and by linear structures such as ditches, brooks, road field paths and terraces. Forested areas, settlements and flat valley bottoms are not taken into consideration. Soil loss was calculated for each erotop and the erotop map of the catchment was created by GIS aided method.

Table 3 shows the frequency distribution and *Fig. 2* the areal distribution of soil loss values. Most of the values are below 10 t/ha (on 42% of the whole area but on about 90% of the area for which soil loss was calculated). Water discharge and sediment load of the Örvényesi-Séd catchment were measured weekly in Örvényes, near the lakeshore, at the outlet of the stream *(Table 4)*.

Table 3. Soil loss in the Örvényesi-Séd catchment (GeoJournal 1995)

Forest	1200 (ha)	49 (%)
0–1 t/ha	339	14
1–5 t/ha	408	16
5–10 t/ha	287	12
10–15 t/ha	93	4
15–30 t/ha	94	4
>30 t/ha	17	1

If soil loss calculated for the whole catchment is compared with measurement data at Örvényes *(Table 3)* we arrive at the conclusion that only ca. 2% of the calculated soil loss leaves actually the catchment.

Table 4. Sediment yield in the Örvényesi-Séd catchment (t/year) between 1977–1994. Calculations based on sediment load measurements at Örvényes

Year	Sediment (t)	Year	Sediment (t)
1977	68,05	1986	
1978	45,97	1987	173,57
1979	58,97	1988	122,61
1980	188,49	1989	50,46
1981	73,79	1990	13,52
1982	80,16	1991	61,30
1983	177,64	1992	13,93
1984	684,20	1993	20,31
1985	91,45	1994	8,20

According to the above result the relatively small catchments like Örvényesi-Séd on the northern side of the lake do not share a considerable part in sediment influx into the lake therefore further investigations were launched on the southern catchment.

Assessment of soil loss by EPIC

Soil loss was estimated in the northern catchment for the tributary catchment (i.e. that of Örvényesi-Séd)s by the EPIC model (Version 5300). Input data were derived from the available GIS database, from daily data series of Mencshely meteorological station (3 km), as well as from laboratory analysis of 40 soil pits and 179 boreholes. The parameters for various crops and tillage operations had to be adjusted to the local, Hungarian conditions. Data on crop yields were taken from local databases. No data could be obtained for the small privately owned vineyards of the watershed and as a consequence of this they were excluded from the investigation. Thus the model was run on arable land and grassland only. Erotop homogeneity by soil type and land use was established by selecting the predominant category within the units instead of applying any segmentation. Measurement data of soil erosion plots were applied for the calibration of the model.

Comparison of results of USLE and EPIC

The mean sediment yield value calculated by the EPIC model is 12,68 (t/ha/year), on arable land it is 22,14 and on grassland 6,08 t/ha/year. Calculations by the USLE are as follows: 4,18 (mean), 2,35 (arable) and 3.1 (grassland) t/ha/year. The areal distribution of soil loss values calculated for the erotops is presented on *Fig. 3.*

There are, however, great differencies between soil loss values of various soil and land use types *(Table 5).*

A close relationship between slope angle and soil loss values could be shown by the EPIC model is proven by this correlation coefficients on grassland (0,78) and on arable land (0,85) prove this. Applying non-linear regression an equally good correlation in obtained *(Fig. 4).*

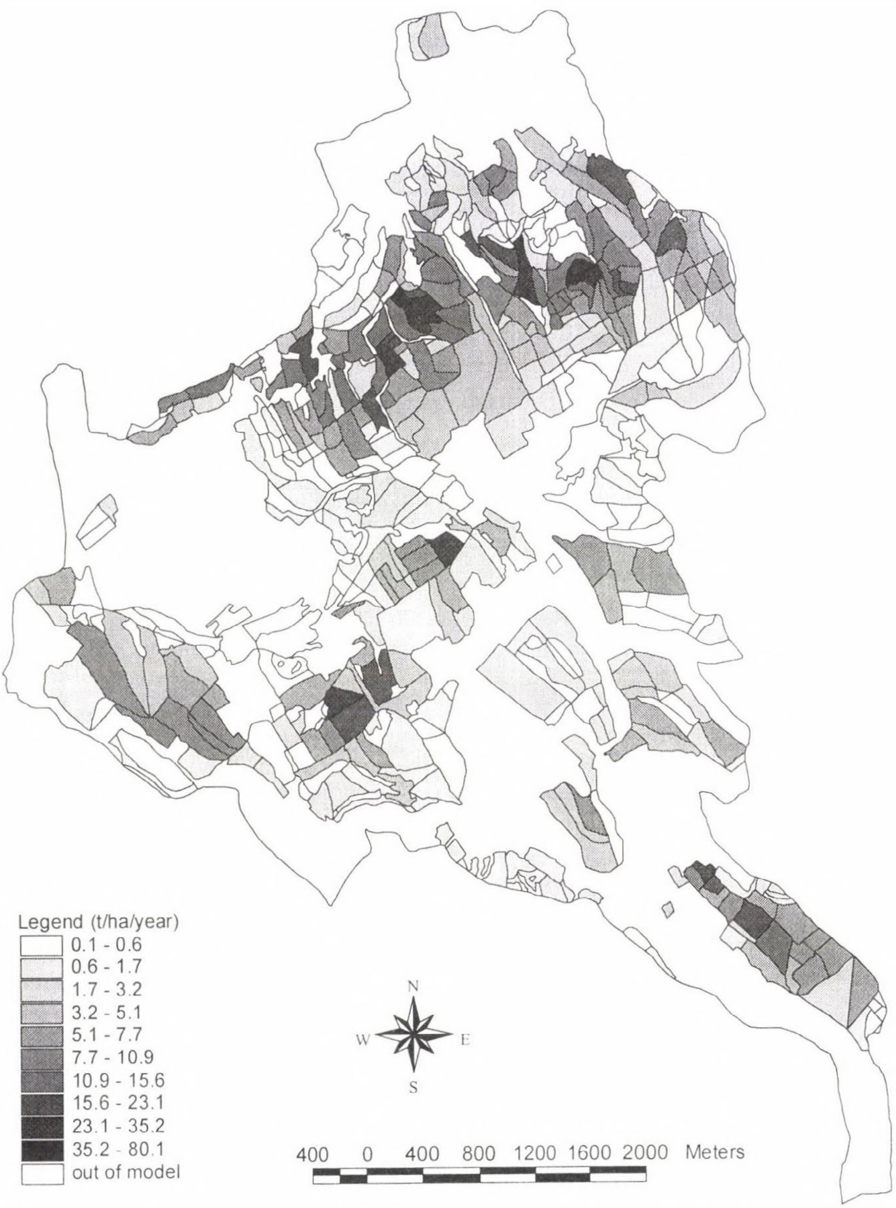

Fig. 2. Soil loss in Örvényesi-Séd catchment calculated by the USLE for erotops

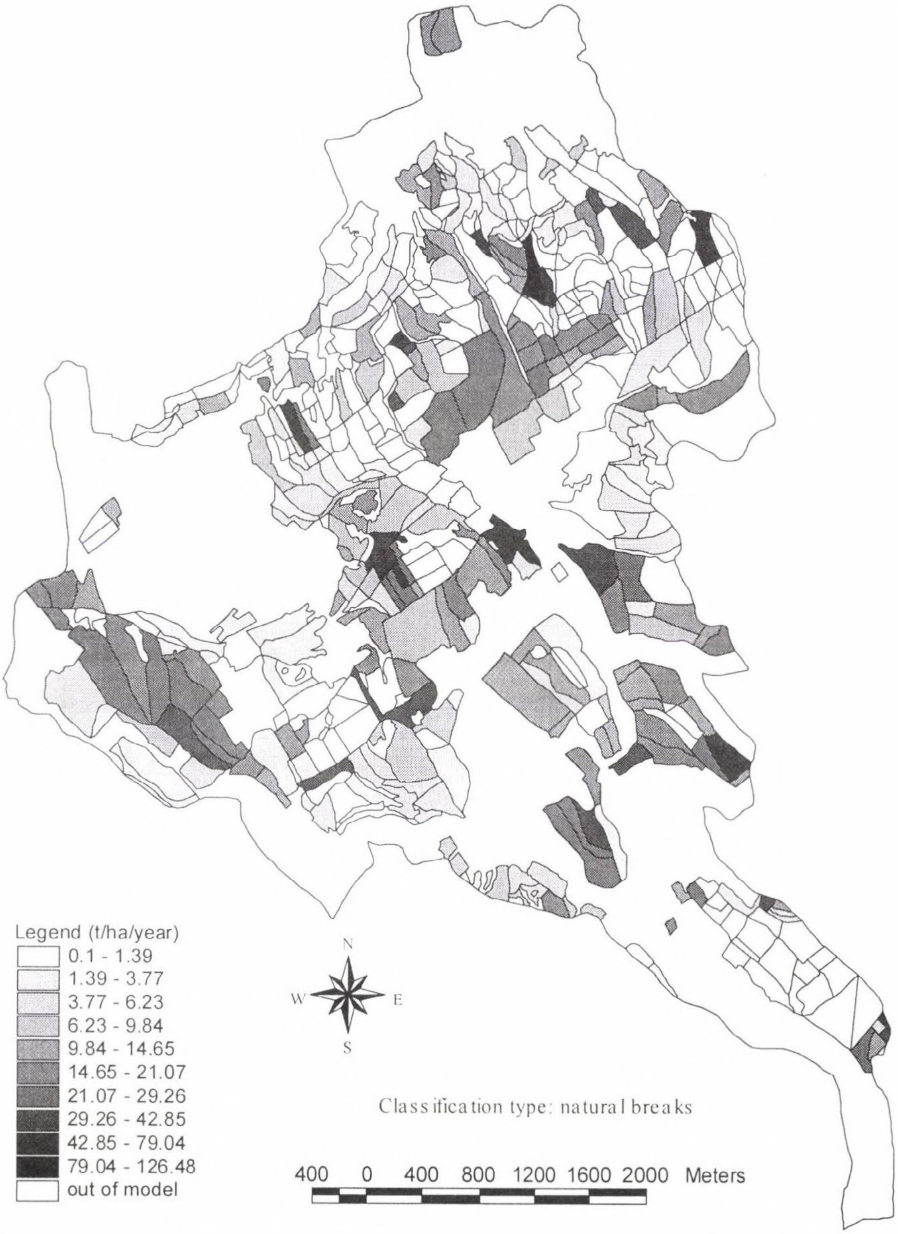

Fig. 3. Soil loss of Örvényesi-Séd catchment calculated by EPIC for erotops

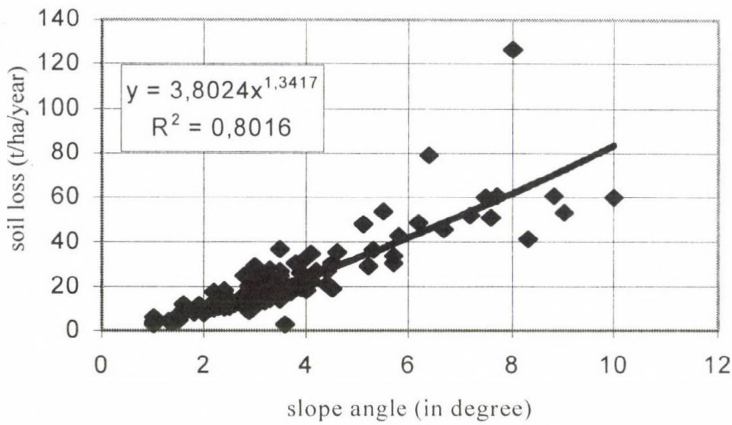

Fig. 4. Relationship between slope angle and soil loss

Table 5. Soil loss calculated for various soil and land use types

Soil type	Skeletal soil	Humus carbonate soil	Brown rendzina	Brown forest soil	Meadow soil	Colluvial soil
Erotops on arable land						
Total area (ha)	41.5	73.7	33.2	199.5	49.4	61.3
Average slope steepness (°)	5.55	3.58	3.05	3.22	3.13	2.85
Mean sediment yield (t/ha/year)	39.89	27.70	15.27	18.83	18.84	18.26
Standard deviation	32.12	14.80	3.21	11.92	17.43	15.15
Erotops on grassland						
Total area (ha)	133.71	37.49	34.91	146.51	16.64	74.13
Average slope steepness (°)	5.89	4.92	4.99	5.41	1.58	2.70
Mean sediment yield (t/ha/year)	6.75	6.41	3.78	7.08	1.11	3.11
Standard deviation	4.94	2.55	3.62	5.51	0.79	2.33

Soil erosion assessment in the Lake Velence catchment

Introduction of the study area

The catchment area of the Lake Velence *(Fig. 1)* covers the south-eastern slopes of the Vértes Mountains, the northern part of the Mezőföld region and the Velence Hills with a total surface of 604.2 km^2. The area is non-uniform, as reflected by the topography, the hilly and flat areas differing both in geological age and structure alike.

The lake is situated in a shallow depression at the foot of the Velence Hills. Its surface is 24.2 km^2 at 160 cm water level. Karstic rocks emerge to the surface on the northern part of the watershed, where most of the precipitation finds access to the deeper formations, so that the runoff from this subcatchment is negligible.

Estimation of soil loss

The Universal Soil Loss Equation (USLE, WISCHMEIER and SMITH, 1978) was applied to estimate soil loss in the catchment, but the method is different from the erotop method presented above. The area was divided into grid cells of 30x30 m. The dominant value of each USLE factor was determined for each grid cell and a map series of the factors were created.

Soil loss in t/ha/year was calculated for each pixel by GIS methods within the frame-work of ARC-INFO i.e. the six maps representing the factors of the USLE were overlain and the RKLSCP multiplication was carried out for each pixel.

The *R factor* (rain erosivity) was identified from rainfall intensity data of Agárd station for the year 1998. The R factor value for 1998 is:

$$R=134 kJm^{-2}\ mmh^{-1}$$

For the calculation of the *K factor* (soil erodibility) values the following soil data are needed: M = silt and very fine sand content; OS = humus content; A = aggregate size; D = soil permeability conditions. Maps for all these subfactors were created and the K factor was calculated by applying the following equation for each grid cell:

$$K = 2{,}77 \cdot 10^{-6} \cdot M^{1.14} \cdot (12\text{-}OS) + 0{,}043\ (A\text{-}2) + 0{,}033 \cdot (4\text{-}D).$$

The M subfactor was defined on the basis of soil texture data from the profiles of the soil monitoring points while OS, A and D were derived from the relating attributes of the 1:100 000 digitised soil map.

The topographic *factors, L* and *S* were derived from the digital elevation model (DEM) using the 1:50 000 topographical map and the Arc/Info hydrology module. The L factor was determined as follows. Using the DEM first the flow direction grid was calculated. Each cell of this raster layer contains the direction of the water flow from the actual cell towards the next cell along the surface of the slope. Based on the flow direction grid the flow length layer i.e. the length of the waterflow from the cell to the watershed was calculated next. Corresponding to the USLE standards the maximal slope length is 1000 m, so the longer flow length values were taken equal to 1000. The following formulas were used to compute the LS values:

$$LS = (l/22{,}1)^{m}(65{,}41 + \sin^2\theta + 4{,}56 \sin\theta + 0{,}065)$$

where:

l [m]= erosive slope length,

θ (°)= slope angle,

$$S = (65{,}41 + \sin^2\theta + 4{,}56 \sin\theta + 0{,}065)$$

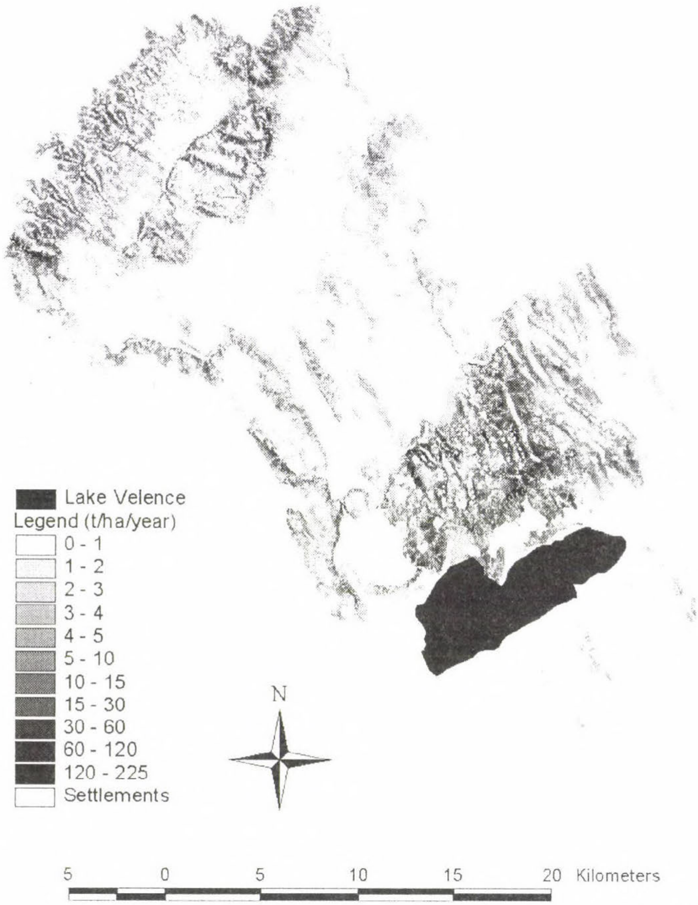

Fig. 5. Estimated soil loss in Lake Velence catchment for 1997–1998.

The *C factor* (land cover and management) was determined using data from the literature for four land use types: forest, arable land, vineyard and grassland.

As no soil protection was applied in the study area, the *P factor* value was 1 for each grid cell.

The soil loss values were calculated for two years. These calculations led to the following conclusions *(Fig. 5)*.

As it could be expected the highest soil loss values and the biggest erosion risk is on the slopes of the mountainous areas i.e. those of Vértes Mountains and Velence Hills. There is, however, a remarkable difference between the two. Velence Hills have much higher soil loss values then Vértes Mountains. The possible explanations follow below.

The morphological situation can explain the difference: slopes are much longer in Velence Mountains and they have also a higher value of dissection. The distribution of genetic soil types contribute to the difference, too. Rendzinas are typical in Vértes Mountains whereas various types of brown forest soils characterise the Velence Hills. Rendzinas generally are less erodable than brown forest soils. On the steep hillslopes of the western part of the investigated area severe erosion can be explained by slope gradient. Flat or slightly undulating areas have of course little or no erosion, in many cases supported by less erodable soil types like different types and subtypes of chernozems. It is conspicuous that stony soils even in hilly areas have very little soil loss. The role of soil texture can be recognised in the silty sand areas of Velence Hills, which are much stronger eroded then other parts of the region. The lower soil loss values of Velence Hills are also explained by their soil texture: they are covered by clayey silt.

Finally it has been established that there is no direct connection between land use and soil loss because of the mosaic-like pattern of the distribution of different land use types. A more detailed survey (1:25 000) would be necessary to find interrelationships if there are any.

CONCLUSION

The paper presented three case studies based on two models. Two different GIS solutions of the USLE were presented, i.e. a vector and a raster based method. Both are well applicable having advantages and disadvantages. The first case study was intended to give long term average values of soil loss, while the second one was elaborated only for two years. Because of the considerable differences of the R factor the latter is suggested to be applied in the future. The EPIC model is also well applicable and the values produced by it do not differ very much from those calculated by the USLE. The problem is that the former is more difficult to be parametrised than the latter.

The two models used in this study were elaborated in the USA based on research, measurements and experiments in that country. Their application somewhere else should therefore be preceded by a careful study of model parameters and model validation is extremely important. The applications presented above are all based on very careful parametrisation and validation. But even so they are not as reliable as a model to be elaborated for Hungarian conditions. The other problem is that they do not really follow the process of erosion, as it is well known. This statement concerns first of all the USLE. They both are more or less black box models for the applier. The main advantage is that they have been used in many countries of the world. The best way would be, however to build a special model for Hungarian conditions.

Acknowledgement – The national projects were funded by the National Science Foundation (OTKA), project numbers (T 001277, T 024165). Of the two international projects Balaton project was supported by the German Research Fund (Deutsche Forschungsgemeinschaft) and by the Hungarian Academy of Sciences, and Lake Velence project was subsidised by the Ministries of Environment of Luxembourg and Hungary. All of these supports are gratefully acknowledged.

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NEW ASPECTS OF MAPPING METHODS OF SOIL SENSITIVITY AS A FACTOR OF LANDSCAPE LOADABILITY

SZILÁRD SZABÓ–ATTILA KERÉNYI[1]

Introduction

Chronic soil acidification is a major environmental problem in Hungary. It is a risk to agricultural production and, in an indirect way, to human health. Crops can only grow and provide acceptable yields within certain pH limits. In addition, the reduction of soil reaction may lead to the production of soluble heavy metal compounds, which may be included in the food chain through groundwater and plants.

The process also occurs under natural conditions. The three major factors of acidification, however, all have their origin in human activities: atmospheric deposition, improper application of fertilisers and the disposal of industrial and municipal waste of acidifying effect (BLASKÓ, L. *et al.* 1998).

According to the measurements by the Plant Protection and Agrochemistry Centres of the Ministry of Agriculture between 1977 and 1985, the ratio of samples below pH 6 grew with 7 per cent and the increase of areas with pH below 4.5 was also remarkable (BUZÁS, I. *et al.* 1986).

With multiple regression of acidification values, KRISZTIÁN, J. *et al.* (1995) claim artificial fertilisers responsible for the process to 63.8 per cent, including the 56.5 per cent share of nitrogen fertilisers. In his observations, the portions of nitrogen fertilisers applied parallel with liming resulted in significant increases of yields for several crops. Liming was financed by the government but the system was highly inefficient: exclusively large amounts of lime applied were paid for and, therefore, the state or co-operative farms had to wait until their lands reached a sufficiently low pH and amelioration could be started only then (VÁRALLYAY, GY. 1994). Today, however, there is no support of any kind and liming is the privilege of some better-off farmers.

As a consequence, authors placed soil acidification into the focus of their investigations since it is a process which also affects landscape functioning (changes in seminatural vegetation, crop yields). The objective was to consider the necessity of weighting through the comparison of results from unweighted and weighted approaches.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 026318.

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Methods

The Department of Applied Landscape Geography at the University of Debrecen has studied the area of Bogács on the Bükk foothills (*Fig. 1*) for several years. Thus, laboratory analyses of 330 samples from 110 soil profiles prepared in the area between 1979 and 1988 were available for the investigation. The data were complemented with the determination of several special parameters used in environmental studies and fed into a digital data base. Finally, in order to check the reliability of the soil acidification map received, further data from the analyses of soil samples from the period 1997 to 1999 were added.

The soil profile data were grouped according to depth: 0 to 25 cm; 25 to 50 cm and the deepest point of the profile. The reason behind the application of this procedure instead of identifying genetic soil horizons is that soil profiles had various depths and no genetic horizons of equal depth could be determined but this approach allowed the separation of two horizons of the same depth (25 cm). The points of maximum depth in each profile inform about tendencies occurring towards greater depths.

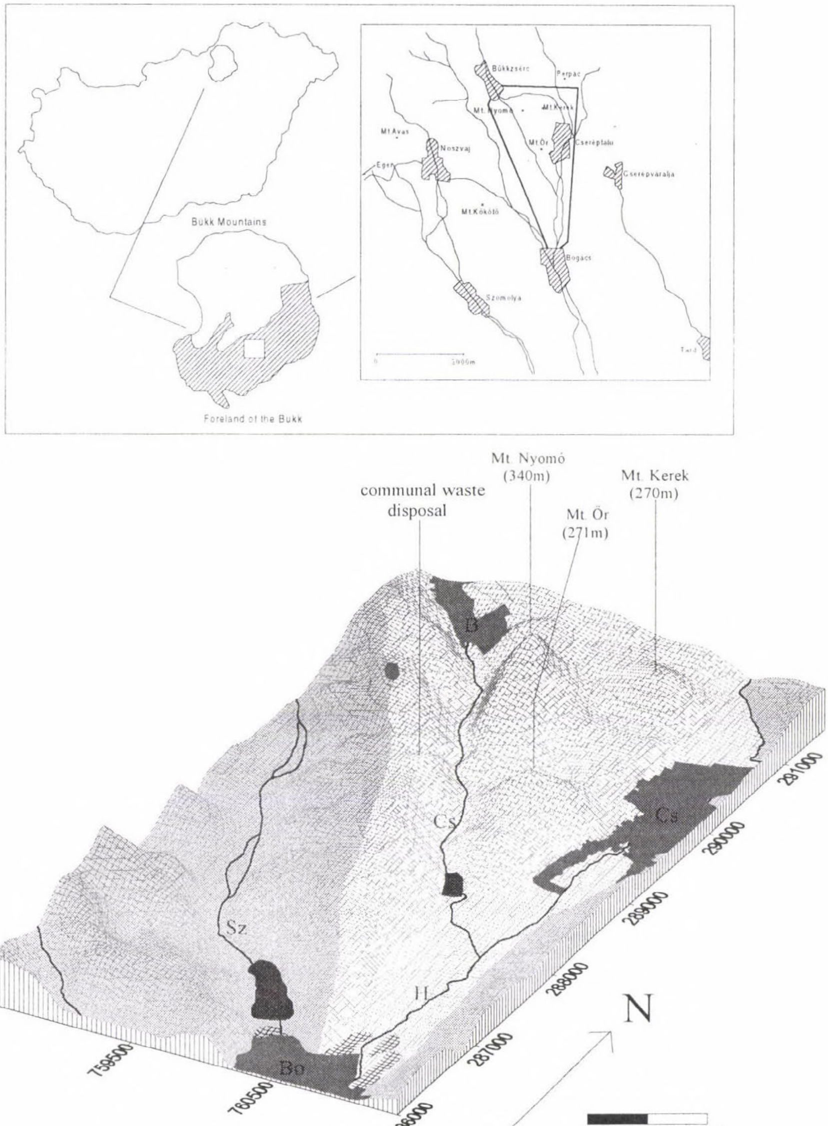
The map of soil sensitivity for acidification was produced through the weighted addition of data describing the three soil layers. The parameters included in the investigation were the following: pH, hydrolithic acidity, clay content, $CaCO_3$ content, amount and quality of humus, depth of humous horizon and buffering capacity.

In weighting the indirect and direct indicators like soil reaction (KCl), humus and clay contents received a factor of 1, hydrolithic acidity (y_1), $CaCO_3$ content, buffering capacity and the special environmental capacity of humus content by HARGITAI, L. (1983) a factor of 2 on the basis of contribution to acidification. Varying in dimensions, the data had to be transformed first into percentages.

Mapping was done by minimum curvature interpolation technique using Surfer For Windows 6.04. The outcomes were further processed using Idrisi For Windows 2.0 as this software includes a mapping algebra which allows the multiplication of values in the pixels of the map (SÁRKÖZI, F. 1996; TAMÁS, J. and DIÓSZEGI, A. 1996) and their addition during weighting instead of calculating for the sampling sites. The reduction in the number of samples involved in the latter method can be eliminated (since only those profiles can be regarded where all parameters are available for all layers of consideration, it is only possible to add them up in this case).

Maps were first integrated layer by layer and then the corresponding data for each layer were added up. The final product was a map which shows areas of various degree of hazard (*Fig. 2*).

The summing was performed without weighting (*Fig. 3*) and the maps were compared to see the changes observed in sensitivity if the mentioned soil properties are investigated separately. Finally, the option was also studied how reliable the results are if only the uppermost 25 cm of soils is analysed with weighting (*Fig. 4*) or without that (*Fig. 5*) since it is not always possible to dig soil pits (for lack of resources or time). For the comparison Idrisi 2.0 cross-tabulation was used again, which allows the analysis of data by pairs of maps. This Idrisi 2.0 option provides two values for similar-



B: Bükkzsérc
Cs: Cserépfalu
Bo: Bogács

H: Stream Hór
Cs: Stream Cseresznyés
Sz: Stream Szoros

Fig. 1. The test area

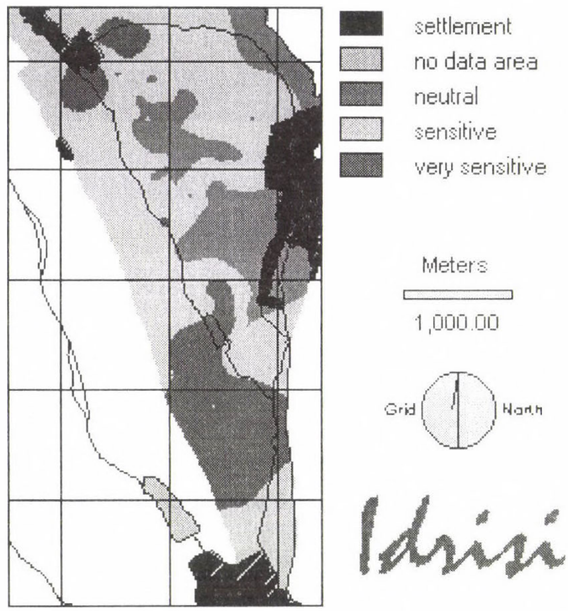


Fig. 2. Soil sensitivity cartogram (total profile, weighting method)

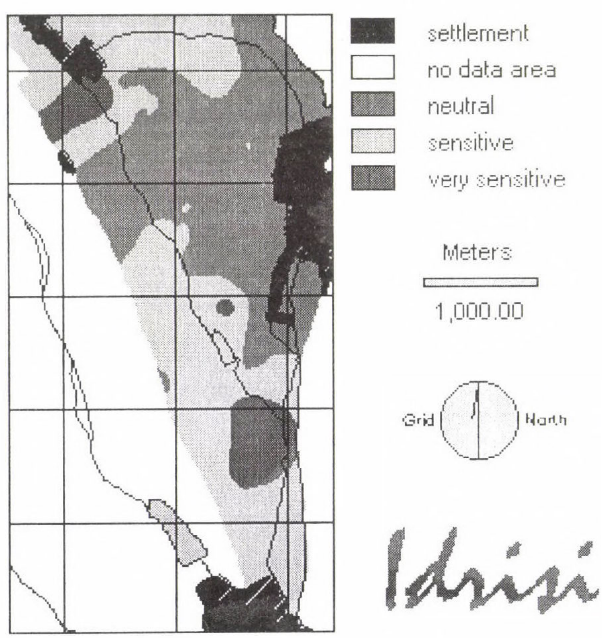


Fig. 3. Soil sensitivity cartogram (total profile, without weighting)

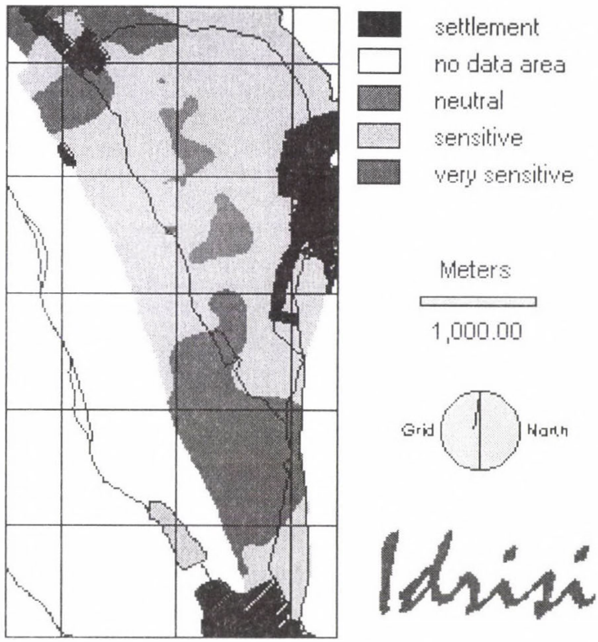


Fig. 4. Soil sensitivity cartogram (uppermost 25 cm layer, weighting method)

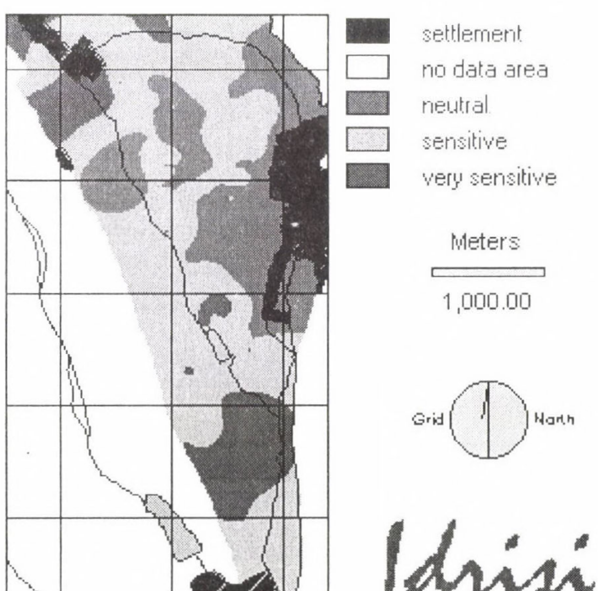


Fig. 5. Soil sensitivity cartogram (uppermost 25 cm layer, without weighting)

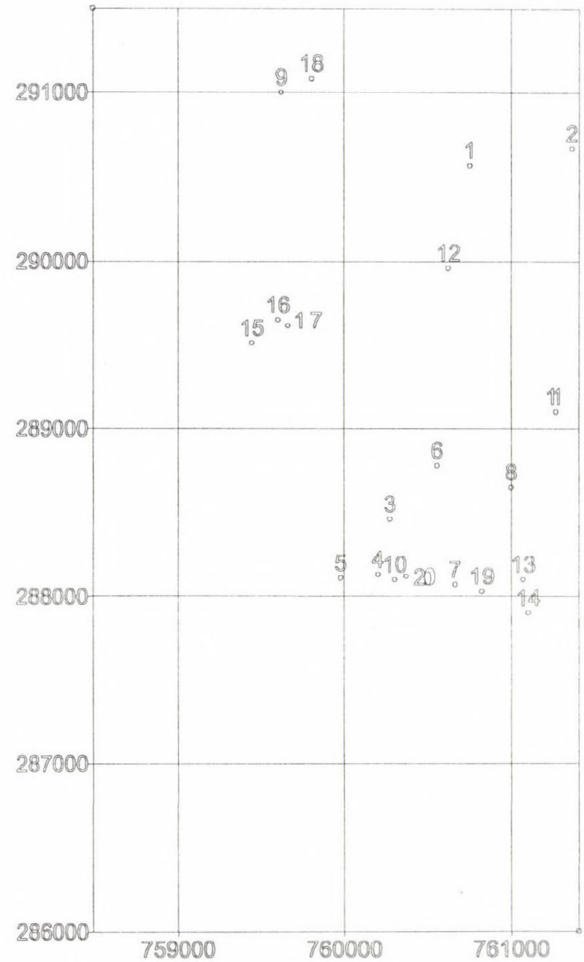


Fig. 6. Coordinates of recent (1997–99) samples

ity: Cramer's V value and the Kappa Coefficient Index of Agreement. The two indices investigate two pictures whether their differences are incidental or reflect actual variation. They range from 0 (total incidentality) to 1 (perfect and true agreement). A value of 0.14, for instance, can rather be regarded as 14 per cent better than entirely incidental (SÁRKÖZI F. 1996).

The accuracy of the map received was checked by the soil samples taken from 1997–99 (*Fig. 6*). For this task the parameters pH (KCl), y_1 and y_2 are used as indicators of potential acidity. The correctness of the model was estimated for a total of 20 samples. Since the 20 samples were not sufficient for producing a new map showing areal changes, another method of checking had to be employed. Relying on papers by STEFANOVITS, P. (1977) and KÁDÁR, I. (1998), values of acidity ranging from 1 to 3 were identified in the manner described below

– where pH(KCl) is above 5.51 and y_1 is below 4 and acidity is low or negligible (neutral class) – acidity is 1;

– where pH(KCl) is 4.51 to 5.5, hydrolithic acidity is 4 to 8 and acidity is medium (sensitive class) – acidity is 2;

– and finally, where pH(KCl) is below 4.5, y_1 is above 8 and acidity is high (very sensitive class) – acidity is 3.

Out of the 20 samples four were taken in the same site as before, using the same sampling technique. In four cases the samples, while in the rest of the sites the correctness of the acidity map were tested.

Results and conclusions

The research findings can be interpreted in two possible ways. The comparison by cross-tabulation provides preliminary information on the differences between the maps created. The checking on the basis of soil samples evaluates the summing techniques employed. *Table 1* presents the results of comparison between the four maps of sensitivity to acidification.

Table 1. Similarity matrix of maps (the upper value is Cramer's V value, the lower is the Kappa Index)

Map	Complete profile, with weighting	Complete profile, without weighting	Uppermost 25 cm only, with weighting	Uppermost 25 cm only, without weighting
Complete profile, with weighting	–	0.7606 0.7703	0.7340 0.7099	0.7523 0.7814
Complete profile, without weighting	–	–	0.6938 0.5139	0.7095 0.6648
Uppermost 25 cm only, with weighting	–	–	–	0.8077 0.8157
Uppermost 25 cm only, without weighting	–	–	–	–

It is obvious that the agreement between maps ranges from 50 to 80 per cent and the similarity indices are invariably above 0.5. The logical conclusion could be that the laboratory analysis and evaluation of surface samples is sufficient. This statement, however, can only be accepted with reservation since the high values suggest that areas outside the test field, which could not be left out from the comparison, represent a considerable proportion (42.29 per cent) of map pairs. Therefore, agreement was calculated for classes and the outcome is summarised in *Table 2*. Here it is clear that there are major deviations just in the cases of the most endangered areas. While there is an agreement of 98 to 100 per cent in some instances, in others it decreases to 68–69 per cent or even drops to 28 per cent.

Through the areal analysis of maps a comprehensive picture is gained on the sensitivity of the study area to acidification. *Table 3* shows the distribution of the individual sensitivity classes by area.

The maps integrated in various ways provide different results as far as the extension of sensitivity classes by area is concerned. It is shown that the proportion of areas assessed as resistant with certainty remains around or under 50 per cent in all cases. In order to select the most appropriate technique, the recent soil samples were examined *(Fig. 6)*.

Samples from the same sites show an unambiguous reduction of pH (–0.12 to –0.8 change in pH).

Table 2. Result of paired comparison of classes on maps of sensitivity to acidification, per cent (deviations compared to classes in columns)

Map	With weighting, complete profile	With weighting, uppermost 25 cm	Without weighting, complete profile	Without weighting, uppermost 25 cm
with weighting, complete profile	–	50.64 74.88 87.19	100 72.98 68.16	64.16 79.80 69.08
with weighting, uppermost 25 cm	99.83 59.68 36.88	–	100 40 28.8	66.68 83.97 100
without weighting, complete profile	17.19 85.47 98.64	8.54 59.2 98.6	–	11.96 76.29 92.74
without weighting, uppermost 25 cm	86.53 73.68 57.11	100.00 72.49 51.17	93.42 60.16 52.97	–

Table 3. Distributions of sensitivity classes on maps by area (ha) and percentage in the study areas

		With weighting, complete profile		With weighting, uppermost 25 cm		Without weighting, complete profile		Without weighting, uppermost 25 cm	
1	No to low sensitivity	196.71	22.2	397.73	45.0	33.89	3.8	481.78	54.5
2	Sensitive	521.9	59.0	415.97	47.0	610.97	69.1	265.31	30.0
3	Very sensitive	164.71	18.6	69.62	8.0	238.46	26.9	136.22	15.4
	Total	*883.32*	*100.0*	*883.32*	*100.0*	*883.32*	*100.0*	*883.32*	*100.0*

The results of checking are summarised in *Table 4*. The number of control samples is not sufficient for drawing final conclusions but some observations can be made.

– Highest precision (70 per cent) is acquired by the map of weighted integration representing the complete profile. This points to the need for weighting but, at the same time, emphasises the need for the refinement of the method of weighting.

– In other instances the agreement remains below 50 per cent.

– The result of testing could be improved by increasing the number of control samples.

The paper was meant to draw attention to the insufficiency of simple superimposing of map layers. A much more reliable picture can be gained through applying parameters with proper multipliers and summing the layers. The resulting final map of sensitivity to acidification may be an appropriate tool for the identification of areas with the most serious hazard of the reduction of pH and for estimating the tendency.

Table 4. Results of control investigation of sensitivity maps and indices of potential acidity of soil samples

No	st	snt	sf	snf	Value of sensitivity	PH (KCl)	$CaCO_3$ (%)	y_1	y_2
1	2	2	1	2	3	6.12		9.06	0.26
2	1	2	1	1	1	7.2	3.19	0	0
3	3	3	2	2	2	4.7		0	0
4	3	3	2	2	3	5.42		9.2	0
5	3	3	2	2	3	5.64		10.25	0
6	3	3	2	2	3	5.03		14.73	0.26
7	3	3	2	3	3	5.27		8.94	0
8	2	2	1	2	2	5.6		5.87	0
9	3	2	2	2	3	4.83		13.94	0.26
10	3	3	2	2	2	6.08		7.89	0
11	1	2	1	1	1	6.21	0.97	0	0
12	2	2	1	1	3	5.14		12.36	0
13	2	2	2	2	2	6.23		6.84	0
14	3	3	2	3	3	4.96		17.1	0.53
15	2	2	1	2	2	5.96		7.89	0
16	2	2	1	2	2	6.42		6.84	0
17	2	2	1	2	1	7.5	1.92	0	0
18	3	2	2	2	3	4.24		17.68	0.36
19	2	3	3	3	2	4.94		3.87	0
20	1	3	2	2	3	4.12		22.52	0.48
agreement	14	10	6	10					

st = complete profile with weighting; snt = complete profile without weighting, sf =. upper 25 cm with weighting; snf = upper 25 cm without weighting.

In the future stages of research the correctness of the acidity model will be refined in the above described way. In addition, land use as a decisive factor is planned to be involved in the assessment as the given crops and, closely related to that, cultivation and land management may substantially modify the process.

Acknowledgement – The present study was supported by the Hungarian National Science-Foundation (OTKA), research topic No. T026318

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MAPPING VERTICAL RELIEF DISSECTION USING GIS

RICHÁRD KISS–GÁBOR MEZŐSI[1]

Introduction

Studies on erosion is a traditional subject in geomorphology. In the last 20 years the geomorphology has moved away from the investigation of the denudation chronology towards the study of processes (EVANS, I. 1998). The traditional problem of geomorphology is the relationship between forms and processes. To explain the landforms and to calculate the characteristics of the surface need a quantitative description of the relief. In the 1970's the morphometry provided a solution for the problem of quantification of the topography (STRAHLER, A.N. 1968; ZEVENBERGEN, L.W. and THORNE, C.R. 1987; MOORE, I.D. *et al.*, 1991). Since the middle of the 1980′s GIS have proven to be a very powerful and useful tool offering advantages for the establishment of spatial distribution of geomorphological processes (MONTGOMERY, D. and DIETRICH W.). The surface itself reflects the potential response to the impact of past exogeneous processes.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 029246.

Background and concept

In this project an attempt was made to find out the constraints of GIS in morphometrical analysis. The aim is not only to produce and analyse a map of vertical relief dissection, but to point out the regional differences in erosion and the translocation of the latter in a geological scale. According to our idea the map of vertical relief dissection (i.e. the negative relict surface) can be constructed for each stream order by subtracting the real surface from the summit planes fitting on the watersheds (CHURCH, M. 1992). Such maps can give information about long-term translocations of erosion and about the changes in its rate. The mapping of vertical relief dissection was based on morphometrical analyses during the 1970′s, but owing to the lack of suitable methods it had been carried out only in limited conditions. The surface modelling offered by GIS have opened new prospects in this direction.

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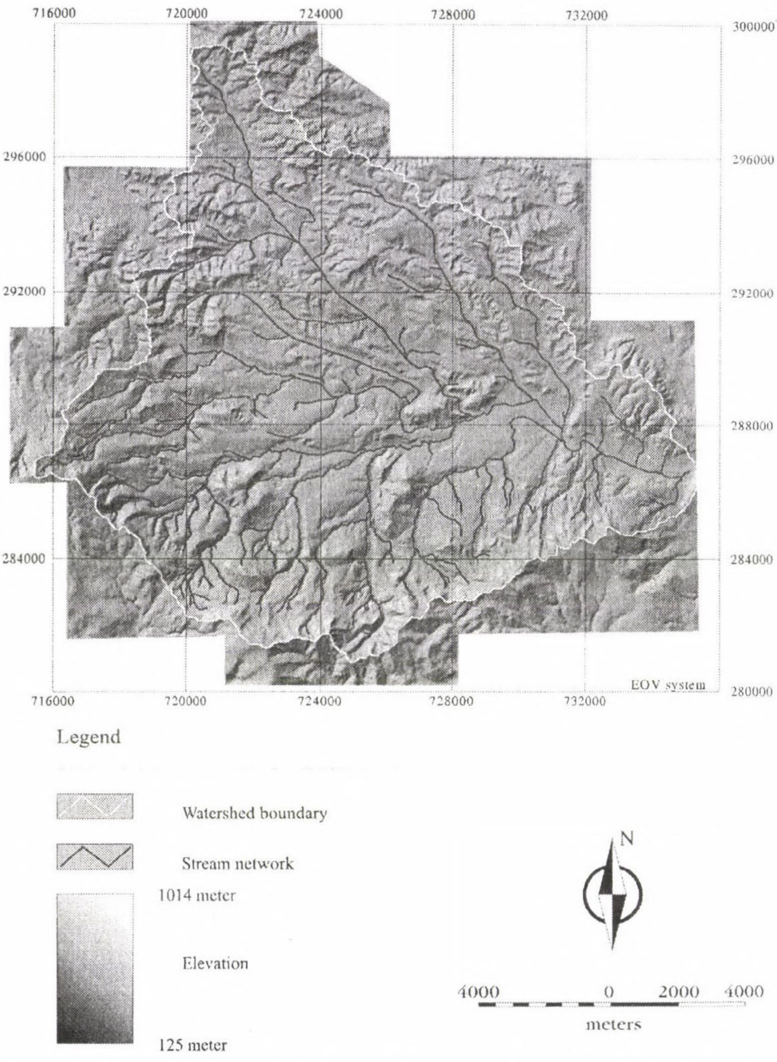

Fig. 1. The orography and drainage system of the watershed of Parádi-Tarna Stream

Method

The analyses were carried out within a catchment of approximately 100 km^2 area (*Fig. 1*). It is situated on the north-eastern edge of the 15 million year old Mátra volcano (North Hungary), mostly built up of andesite and rhyolite tufas. The test area being more or less homogeneous geologically, it was supposed that the direction of the

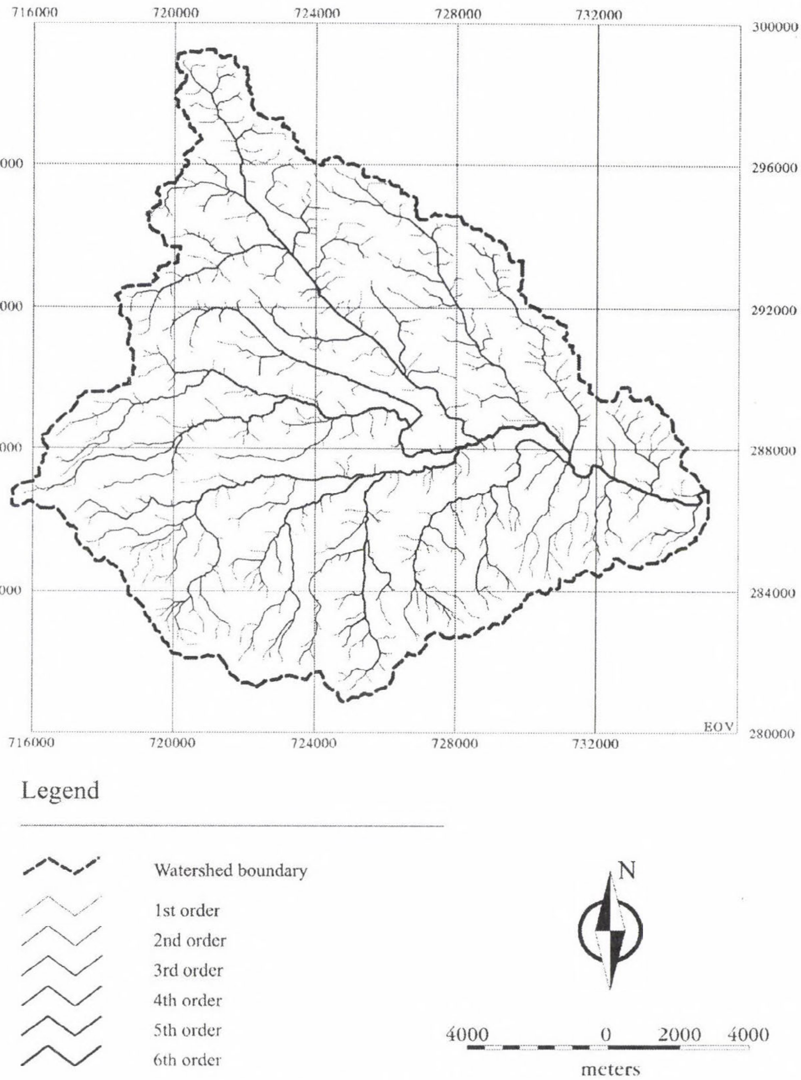

Fig. 2. Strahler's order of the streams

streams had not changed considerably. Besides, there are uniform precipitation conditions, so no significant change in the rate of erosion was expected.

It was supposed that a currently six-ordered stream (according to STRAHLER) had been five-ordered previously, four-ordered even earlier, and so on. This is true only as a model, because despite of a general homogeneity, geological, climatological and orographical disturbances might occur. The suitability of the order system for statistical/GIS investigation was also supposed. The aim of the research is to highlight these disturbances, which can be investigated in the drainage network. First of all, the stream network of the area was categorised following STRAHLER's system (COSTA-CABRAL, M.C. *et al.*, 1994; FREEMAN, T.G. 1991; *Fig. 2*). More than 1000 first-ordered streams were counted. The large number of streams confirmed the statistical results of the drainage analysis. Secondly, the real surface heights were subtracted from the summit planes fitting on the watersheds of first, second, etc. order. In this way vertical relief dissection maps were obtained for each order using ArcInfo 7.0.3. surface modeller. As a preparation a digital terrain model of the test area was compiled.

Results

The maps of vertical relief dissection belonging to the different orders can be seen on *figs 3* through *5*. Their analyses show that two „proto"-rivers started to dissect the surface from south to north and from north-west to south-east, following the main slope directions. One was formed on the southern part of the catchment on Miocene andesite and rhyolite lava, an other one has cut into a loose Oligo-Miocene sandy material (*Fig. 6*). The sinking and swaying of the Mátra Mountains resulted in regional differences of the rate of erosion: on the southern part small basins were formed bordered by dykes, and the relative relief had not increased significantly; in the northern part the relative relief grew considerably, so here the erosional intensity became more accentuated than on its southern counterpart (*Fig. 5*).

Another important change in fluvial erosion could happen during the Pliocene, when the central part of the Mátra Mountains uplifted by about 150–200 m. Therefore, the rate of erosion increased in the southern part of the catchment being closer to the centre than to the northern part.

It is well known that there is an exponential relationship between the stream orders and the total length of certain stream orders. *Fig. 7* shows that the number of the five-ordered streams (as well as the six-ordered ones) is less than it might have been expected. Their mean length and the volumes (*Fig. 8*) vertical relief dissection are decreasing radically compared to the statistical expectations. It can be explained partly by lithological reasons (the five-ordered rivers situated in the middle of the catchment, in an erosion-resistant environment) and partly by the decreasing relative relief.

The above mentioned geological and orographical influences were studied along the Ilona valley. The *Fig. 9a* shows the vertical relief dissection by different ordered streams of that valley, which runs from south to north and than turns to east. The disturbance of the curve of the vertical relief dissection, for example that belonging to the five-ordered stream – at 13,350 m far from its source –, can be explained by geological reasons: here the valleys reached the lower lying lava layers, therefore the drainage pattern became sparser. The regular rhythm of the curves can be plotted as semicircles *(Fig. 9b)*. The integration of the changes in rates of erosion into the exact geological time-scale has not been solved yet.

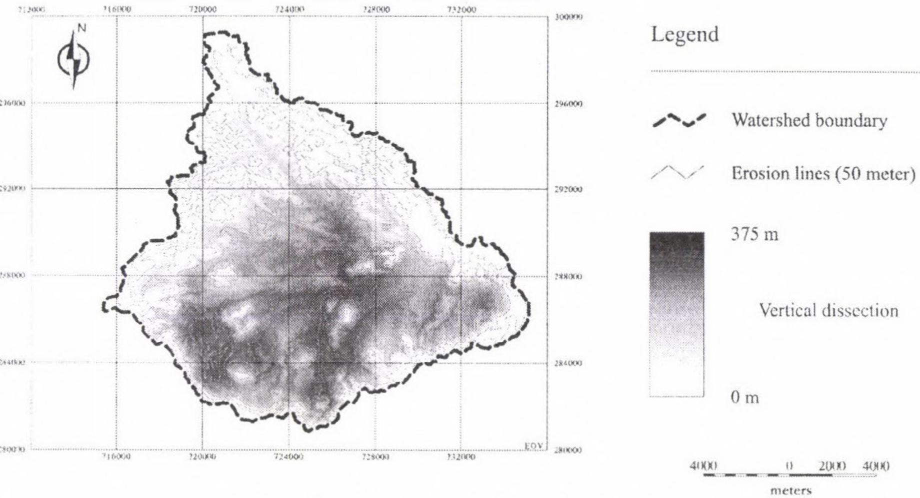

Fig. 3. 6^{th} order relief-dissection of the test area

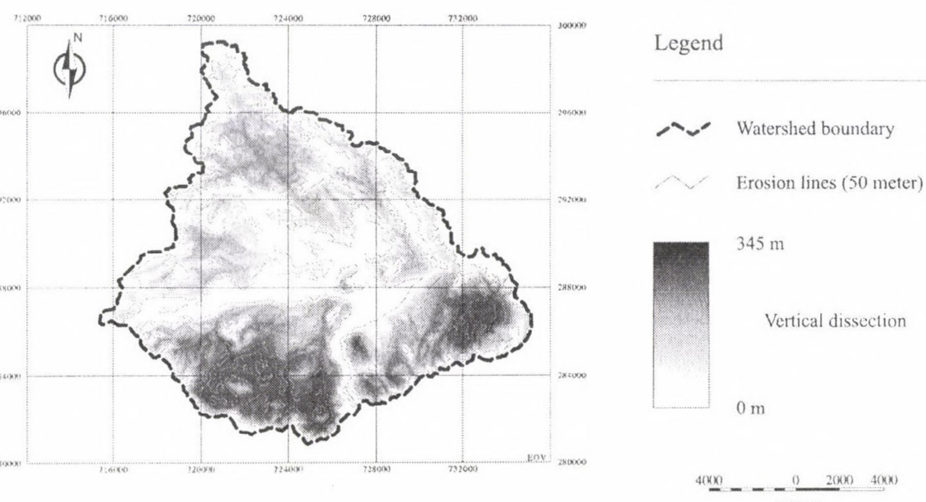

Fig. 4. 5^{th} order relief-dissection of the test area

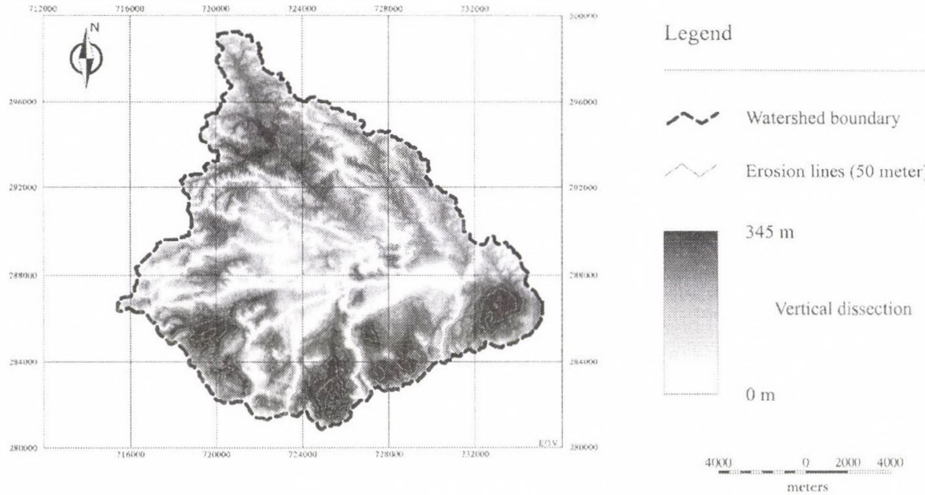

Fig. 5. 4^{th} order relief dissection of the test area

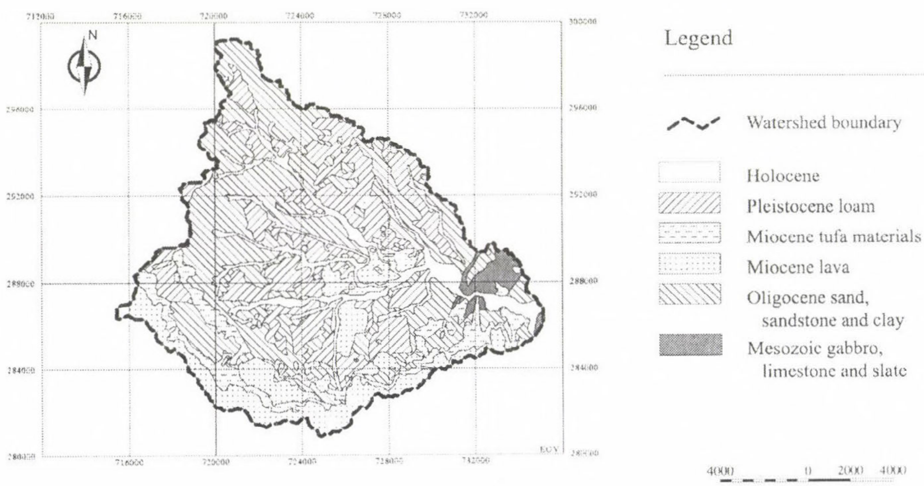

Fig. 6. Geological map of the test area

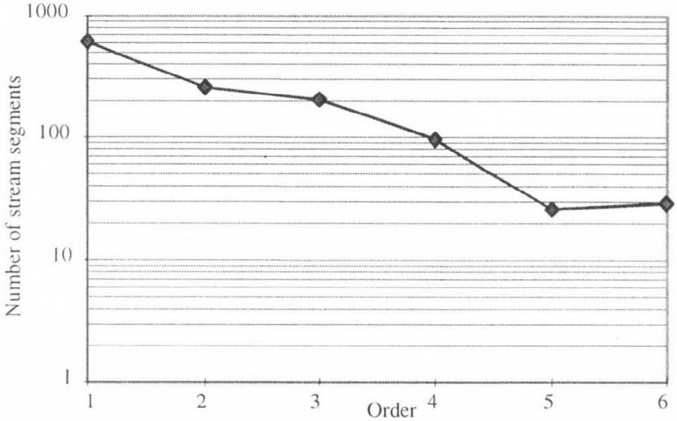

Fig. 7. Frequency of the stream segments

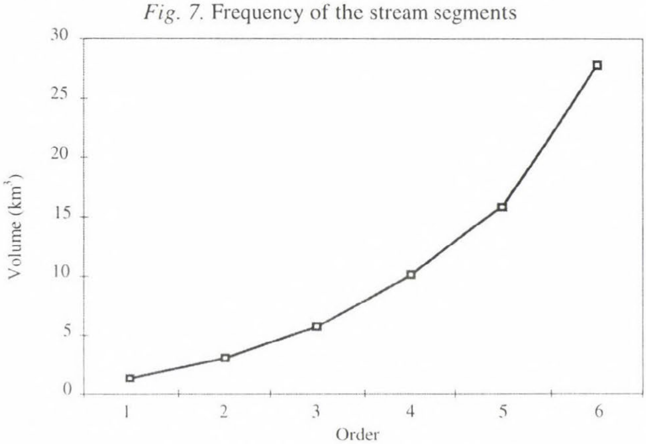

Fig. 8. Volume of the relief dissection

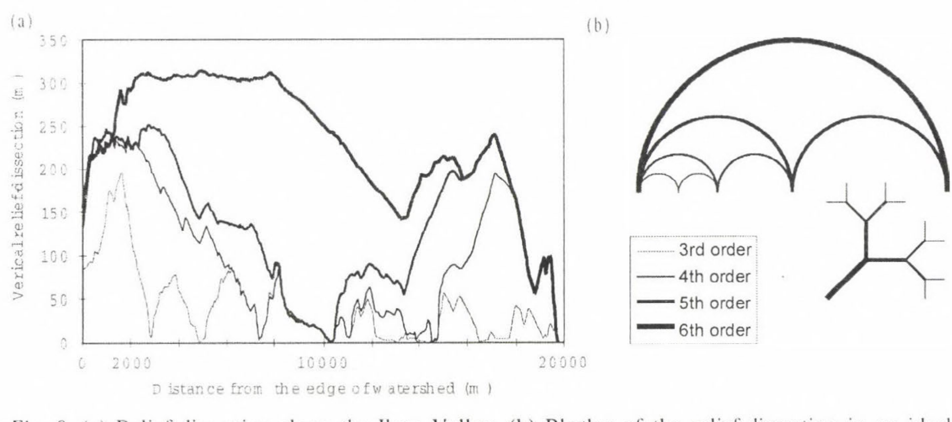

Fig. 9. (a) Relief-dissection along the Ilona Valley. (b) Rhythm of the relief-dissection in an ideal network

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KARST IS A BIOLOGICAL PRODUCT

LÁSZLÓ JAKUCS[1]

Traditional explanation of the karst process

It is not an unknown phenomenon in the history of the natural sciences that axioms which are out of date and have been disproved stubbornly continue to survive for a long period in textbooks, encyclopaedias and even in handbooks. And, although the broader new facts have already revolutionised the viewpoint of specialists dealing with the topic in concern at a research level, the earlier scientific „belief" that has become outdated prevails in the public opinion for a long time. This is precisely what has happened in recent times in connection with the interpretation of karstification and karst phenomena.

Even at present, the traditional textbook scheme interprets karst *phenomena as the rock-dissolving action of precipitation water*. However, as regards the essence of the matter this is erroneous. It has been proved that rainwater in itself has *scarcely any* dissolving action on limestone! The very weak limestone-dissolving activity of surface water resulting from snow and rain would in itself never be sufficient to give rise to the great variety of karst phenomena. In contrast with this, modern science has unambiguously demonstrated that most of the karst phenomena on the surface of the Earth reflect the effect of *the activity of the biota*. Indeed, it has also been proved that even in some of the subsurface karst phenomena, such as cave dripstone formation, the most important transmitter of the genetic process is the biological factor. Karstification is thus a characteristic and exlusive feature of the Earth in our solar system, the extent and nature of the process being strictly proportional to the biological activity of the surface vegetation and the soil.

The generally-known traditional interpretation of karsts was developed centuries ago, in the early days of science. The essence of this conception was that the water falling onto bare limestone rocks and permeating through the network of fissures and cracks inside, dissolves limestone by acting as weak carbonic acid, formed from the carbon dioxide brought with it *from the atmosphere*. As a consequence of the dissolution, the surface rocks display a special variety of forms; *lapies fields (karrenfelds)* develop, and the permeating water, by dissolving the rock and widening the cracks, causes the limestone to collapse repeatedly, so that cauldrons (bowl-shaped

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depressions), *dolines*, are formed on the plateaus. The water permeating ever downwards in the network of cracks in the limestone combines in the depths; it then continues its enhanced dissolution work to create wide cavities and *cave streams*. Thus, all of the characteristic features of the limestone mass (from the surface dolines to the caverns in the depths) were explained via the rock-dissolving action of the precipitation water.

Recognition of mechanical erosive modell of cave formation

The classical karst theory received the first critical blow when, almost simultaneously on the various continents, research workers began to check the changes in chemical composition of the water permeating into the rock. It then turned out that this water is very quickly converted into a calcium-saturated solution, *at a depth of virtually only a few metres*. On permeating deeper, however, such a calcium-saturated solution is no longer capable (except at best under very special conditions) of dissolving more rock. Accordingly, in the vast majority of cases the water reaching caves at depths of a hundred metres or more is quite inactive as regards dissolution. Instead of carrying out further dissolution, the „karst water" permeating into the depths rather *deposits* the mineral substance transported in solution from above. Dripstones build up from the limestone sediments of millions of falling water drops. This means that *the formation of caves can* in no way be attributed to the dissolving work of karst water permeating into the depths.

It has been proved that the cavity system itself is always created in the interior of the mass of limestone by the flow of some other watercourse, *originating from an external water-catchment*, and primarily by the *erosion* caused by the debris swept into the karst by streams. Hence, the cave is not a product of dissolution, but is a simple erosion streambed under the surface *(Photo 1)*. Caves are therefore *not* necessarily a karst phenomenon, since they are is formed only in karsts which have a watercourse system transporting solid particles of debris from some source outside the karst.

Identification of decisive role of the soil-atmosphere

The final blow to the traditional karst explanation was given by the extensive chemical analysis of the water permeating in. They have demonstrated that the average carbon dioxide content of the global atmosphere is only 0.03%, i.e. it is so low that a litre of precipitation water is not able to dissolve even a milligram of carbon dioxide from the atmosphere. Accordingly, the carbon dioxide taken up from the free atmosphere causes *practically no increase* in the limestone-dissolving ability of chemically pure (distilled) water (only 10–15 mg limestone per litre). If this were the only factor, it is hardly likely that the wonderful dissolution karst phenomena in limestones would have developed on the Earth! The loss of 10–15 mg rock per litre of water is virtually negligibly small. All other rocks (even granite) dissolve to almost the same extent in water.

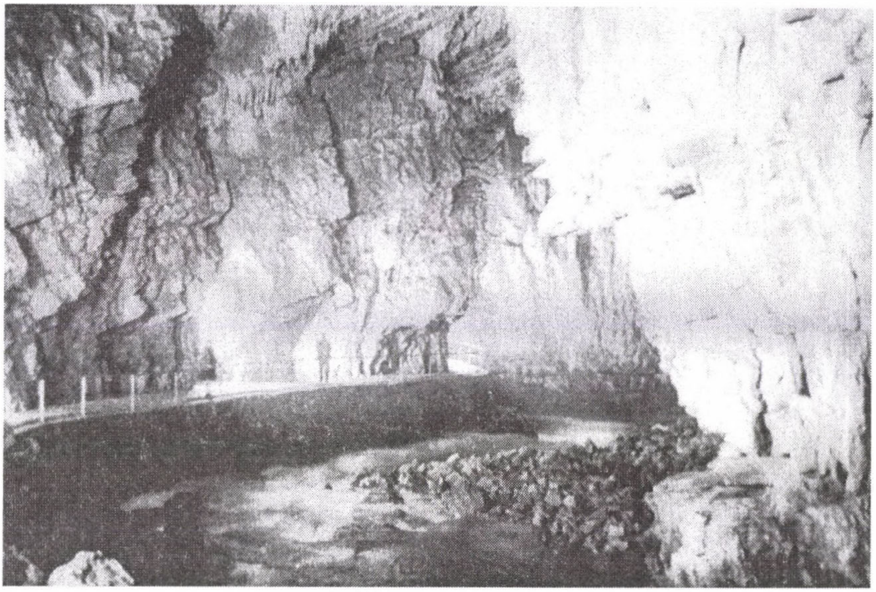

Photo 1. Most cave systems are the products not of dissolution, but of running-water bed-carving erosion. The caves are formed by the flow of water running into the network of cracks in the rock under the surface, in the same way as the mechanism of valley formation may be observed on the surfaces itself. In the genetic sense, therefore, the large cave systems are river valleys eroded under the surface, with all the criteria of bed erosion

Water samples collected from the systems of fissures in the carbonate rocks of karsts, or from the interior of caves themselves, however, show a totally different picture. Their dissolved calcium content may reach even several hundred (sometimes one thousand) milligrams per litre.

Where then does the water acquire such a large amount of carbon dioxide so as to permit it to dissolve a large amount of limestone? In all cases the examinations have clearly indicated this source to be *the soil.* Where the rock is covered by a soil layer, the precipitation must first permeate through this cover before it can reach the rock.

However, in the gas mixture occupying the porous space in soil there is much more carbon dioxide than in the free atmosphere. Here the proportion of this gas is almost always more than 1%, while fairly often it is in excess of 10%. That is, compared to the free atmosphere, at least 30 times, but frequently 300 or more times more carbon dioxide accumulates in the soil atmosphere.

There is no doubt, therefore, that karst water with a high carbonic acid content and with the ability to dissolve much limestone acquires its agressivity not from the air, but from the soil cover. The more the carbon dioxide formed and accumulated in the soil the quicker and the more effective will be the process of the desctructive dissolution of the limestone, i.e. karstification, beneath it.

The carbon dioxide in the soil is produced by the millions of tiny microorganisms living there. This means that the rate of karstification in a given region is controlled not only by the quantity of precipitation permeating in, but even more importantly by *the activity of the biological processes in the soil layer* covering the surface to some depth. That is, the dissolution of the limestone, *karstification, is essentially a formal reflection in the bedrock of the phenomena of biological and chemical development of the pedosphere covering the rock.*

The attractive concepts by the *Cvijić* and *Cholnoky* school is similarly erroneous, therefore. This stated that the reason for the karstification of the *Dinaric Karst* was that, following the devastation of the woods there, the rainwater washed away the soil covering the surface, and the then bared limestone could be freely dissolved by the precipitation. In fact, just the opposite of this argument holds right: the development of the karst phenomena, the corrosion of the dolines and the production of the bizarre rock formations of the lapies, all occurred when the mountains were *covered by woods and soil*. The later baring of the slopes merely revealed all this by making it visible, but at the same time it simultaneously curbed the dynamics of karst development itself.

Naturally, the bioactivity of karst soils is not restricted simply to the carbon dioxide production of the various bacterium and fungus populations living in the soil; the chemical effects of the *roots* under the grasses, bushes and trees living on the soil surface, the decomposition of organic waste, fallen foliage and animal remains rotting in the soil, and many other processes too, may all serve as the sources of carbon dioxide and other acids. Soils with higher bioactivities can be looked on almost as chemical factories, where a huge variety of different *organic acids* are produced. The most important of these are formic acid, acetic acid, oxalic acid, lactic acid, propionic acid, various fulvic and crenic acids, humic and huminic acids, etc. Although the main role is played by carbonic acid, these compounds also take part in the dissolution of limestone to various extents, since they are also dissolved by the water permeating through the soil and transported to the limestone bedrock.

Climatic conditions of the bioactivity of soils and the plant species adequations of karst forms derived from solution

Just as is the situation for the living organisms we know so well directly, the invisible living world of the soils has its own favourable and unfavourable *living conditions*. The biological functions of soil microorganisms react very sensitively to *variations in temperature*, for example. Even the fluctuations in the *daily temperature* are followed closely by a change in the number of bacteria in the soil. Long series of experiments and a large amount of observation material permit the finding, however, that the optimum temperature itself is still not a sufficient condition for stimulation of the population of a soil microorganism; this can be ensured only by the *simultaneous effects of the temperature and soil-moisture optima,* naturally under conditions of satisfactory soil aeration. The decrease or increase of either factor immediately results in a marked decrease in the number of bacteria. Thus, the chemical production of acid in the soil is extremely *climate-sensitive*.

In *tropical soils* with favourable temperatures and moistures therefore, even several hundred times as much carbon dioxide and other organic acids may be formed as in the soils of karsts in the *temperate zones*, for instance. In turn, however, the carbonic acid production in temperate zone karst soils is many times higher than that of the sparse soils covering the cool-surfaced karsts in the *cold zones* and on *high mountains*. It is obvious, therefore, that there are necessarily *tremendous differences* in the intensities of karstification under the different climaties (tropics, desert, Mediterranean, temperate oceanic, high mountains and other cold regions). As a consequence of the climatic sensitivity of the biogenic factors relating to the soil, the aggressivity of water as regards the dissolution of limestone also becomes a function of the climate *(Fig. 1)*. We may be certain that fundamentally it is *these variations* which explain the striking differences in order of magnitude and the very characteristic regional morphologic differences of the karst forms to be seen in various regions of the Earth with different climates.

Under temperate climate, biogenic dissolution is the main genetic factor chiefly shaping the *subsoil lapies* (e.g. root lapies) and the *dolines*. Microorganism populations differing with regard to the species develop in the *rhizosphere*, the root networks of the various plants, grasses, bushes, trees, etc. growing side by side in the soil. As a

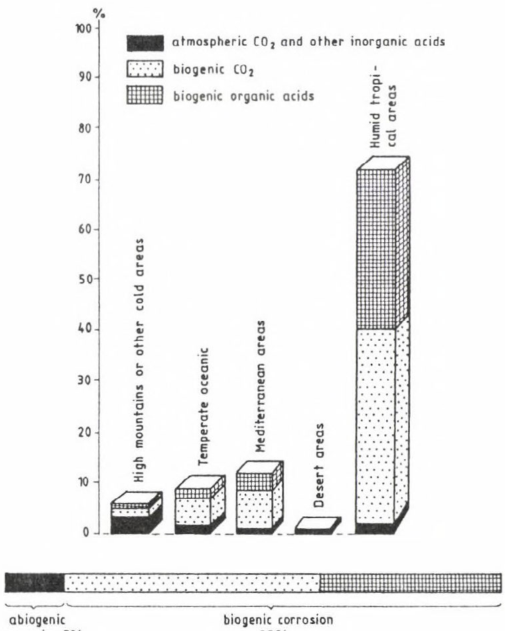

Fig. 1. The magnitudes and factor-proportions of karst corrosion in some characteristic climatic regions of the Earth. The extent of the dynamics of dissolution of limestone is indicated by the height of the columns, and the participation of the factors causing the dissolution processes by the component ratios denoted within the columns

consequence, there will also be qualitative and quantitative differences in the chemical processes in neighbouring rhizospheres or soil regions, and this leads to the variations of acid and gas concentrations in adjacent parts of the soil. The *degrees of aeration* of the individual soil regions depend on the permeability of the soil surface, its moisture content, and exposure, the thickness of the bioactive soil section, and many other factors; this likewise influences the concentrations of liquid and gaseous compounds accumulating in the soil. Hence, even within a few centimetres, extremely great differences may occur in the chemical composition of water permeating through the soil. This *differentiation* in chemical agressivity is in turn reflected in the irregular dissolution forms of the rock, or in the bizarre formation of rock lapies.

Biogenic and abiogenic karren-formations

Bacteria occur in the soil always much denser around the roots than elsewhere. For this reason, in time the roots penetrating into the initially tiny cracks in the rock enlarge these into wider, meandering dissolution channels, which are usually round or oval in cross-section. Limestone riddled with such root channels is *root lapies (Photo 2)*, while extensive rocky surfaces which have lost their soil and become bare are generally known as *lapies fields.*

In the tropics, where both the vegetation cover and the biota concealed by the soil exhibit much more dynamic developments, the effects of biogenic lapies formation are naturally much higher in proportion, too. Here the channels of the root lapies often penetrate the limestone to a depth of even 20–25 m, and the root corrosion may lead to a rock loss by dissolution of as much as 60–70% *(Photo 3)*. The strikingly high intensity of biogenic karstification may be well illustrated by the example of trees making their way through thick limestone layers. In Cuba (but elsewhere in the tropics too), many caves are known where trees have grown through a rock ceiling several metres thick, via chimney channels carved out by themselves *(Photo 4)*.

Otherwise, the karsts not necessarily should become barren; indeed, it is not even a general symptom. However, if this occurs (generally as a result of devastating human impact), the precipitation water will come into direct contact with the limestone of the bared lapies fields on the surface. Trickling down the rock sides, this pure rainwater dissolves the limestone to only a slight extent. Its effect is to produce small dissolution channels, generally parallel to one another and corresponding to the direction of the slopes. This lapies phenomenon, however, is a slowly-developing one, and is not biogenic! The name of this formation is *precipitation lapies or gravitational lapies furrows (Photo 5)*.

Biogenic explanation of karst dolines

It is known by now that *dolines* are also typical biogenic karst forms *(Photo 6)*. They are dish-shaped or cauldron-like depressions, sometimes only a few

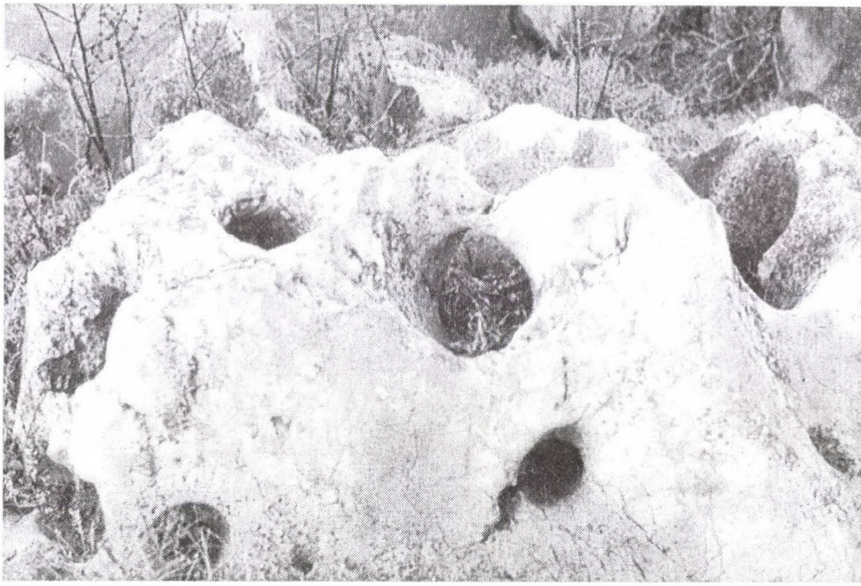

Photo 2. The limestone rock riddled with root channels is evidence of karstification proceeding under the soil. The root lapies to be seen in the photo is a formation of the now barren lapies field above the lake at Aggtelek in Hungary

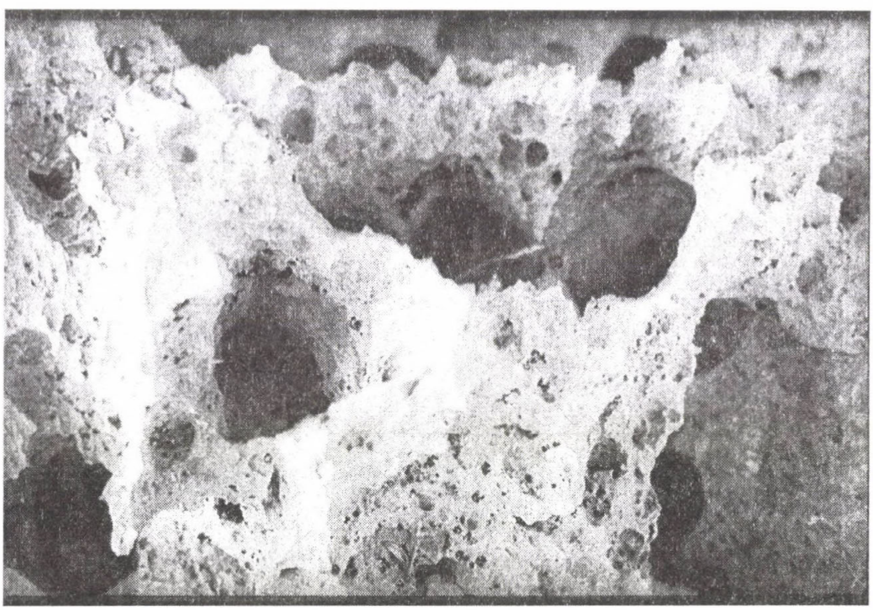

Photo 3. In the abundantly vegetated tropics, root corrosion totally depletes the content of the near-surface limestone layers within a short time. The photo shows a limestone surface in Cuba, where the loss of rock by biogenic dissolution is ca 65–68%

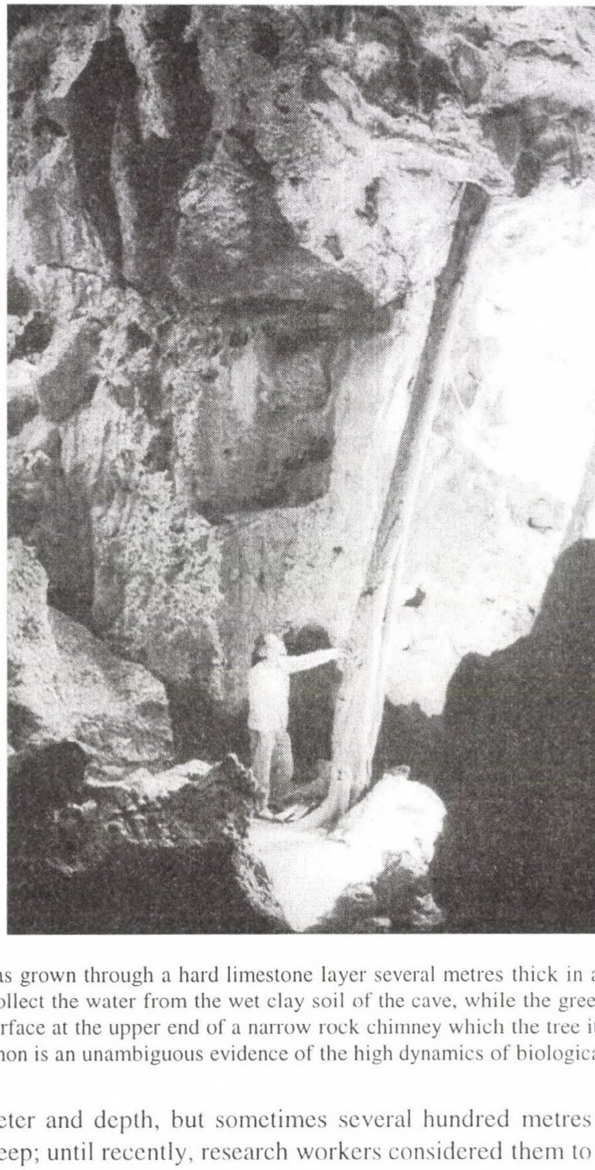

Photo 4. A tree has grown through a hard limestone layer several metres thick in a cave in Cuba. The roots of the tree collect the water from the wet clay soil of the cave, while the green foliage enjoy sthe sunshine on the surface at the upper end of a narrow rock chimney which the tree itself has carved out. This phenomenon is an unambiguous evidence of the high dynamics of biological karst corrosion

metres in diameter and depth, but sometimes several hundred metres across and even 40–60 metres deep; until recently, research workers considered them to be due merely to the collapse of the rock, interpreting them as cave-in phenomena of the underlying caverns and dissolution cavities. It turned out later that dolines and caves do not have much in common. Caves are almost always are situated elsewhere in the depth with no karst depressions located on the surface.

The cave-in origin of dolines also contradicts to the fact that the rock layers on the sides of the dolines almost always retain *their original strike direction and dip*

Photo 5. In contrast with biogenic lapies, the dissolved furrows of abiogenic precipitation lapies formed on bare limestone surfaces exhibit gravitational direction and develop very slowly. In fact the classical karst explanation could give a correct interpretation only for this dissolution form

angle. That is, in the course of the formation of the doline there is no change in the situation of the karstifying layers in which the depression developed *(Fig. 2)*.

The resolution of the contradictory observations and the up-to – date interpretation of doline formation were made possible only by the recognition of biogenic karstification. According to this, a doline is *simply a surface depression caused by dissolution of the rock*, formed at those places on all karst plateaus where the soil covering the rock becomes the most active. Initially, the humus-containing soil particles of loose structure on the higher terrain are easily washed together into flat dissolution depressions, whereby the sites on the karst plateau with an optimum of corrosion begin to be localised into more restricted areas. In time, the solution process (mediated by the soil) is concentrated to an increasing extent in the surface corrosion bowl which begins to develop, since the precipitation is able to wash down the soil ever more effectively

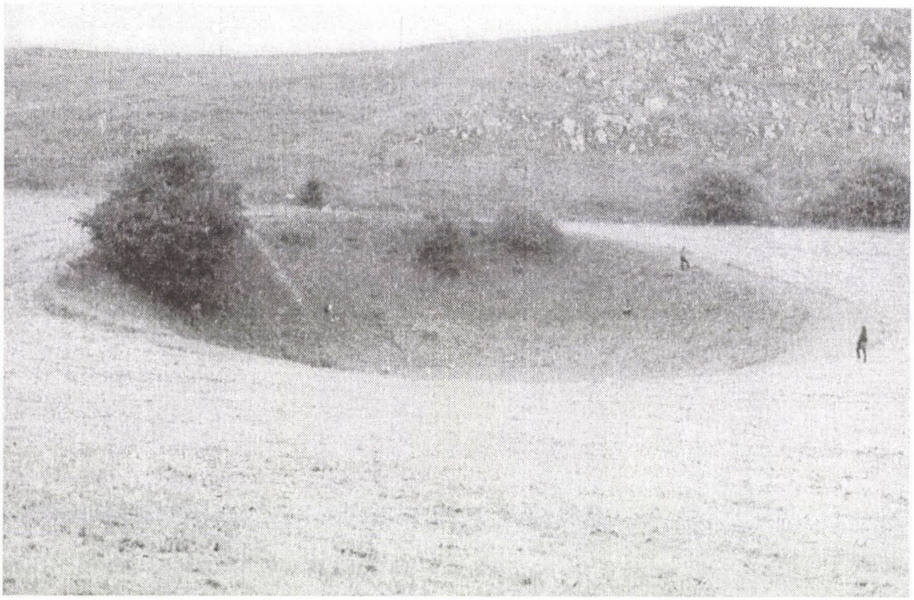

Photo 6. The doline is a characteristic product of biogenic corrosion. The well-known karstic large forms of limestone plateaux are produced by the enhanced rock dissolution typical of the most bioactive soil areas. At the beginning of the process the soil particles from the adjacent surfaces are also washed into the depressions eaten into the still flattish rock; this enhances the areal differentiation in the dynamism of dissolution. By this means, the further deepening of the doline becomes „autocatalytic".

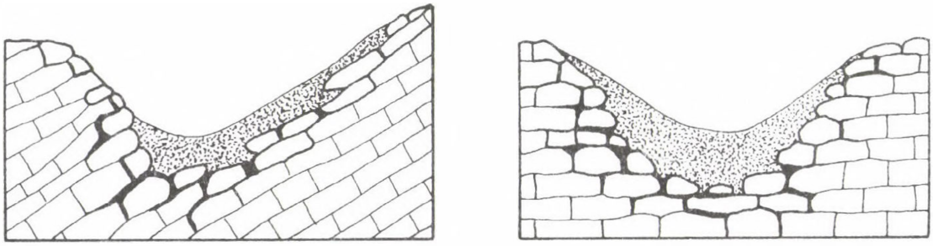

Fig. 2. The profile of dolines „cleaned out" to the standing limestone rocks shows well that the doline formation is not a consequence of the collapse of cavities, but of the local eating-away of the rock surface. The angle of dip of the layers in the doline remains unchanged. The stratification conditions may modify of the base and slope features of the doline

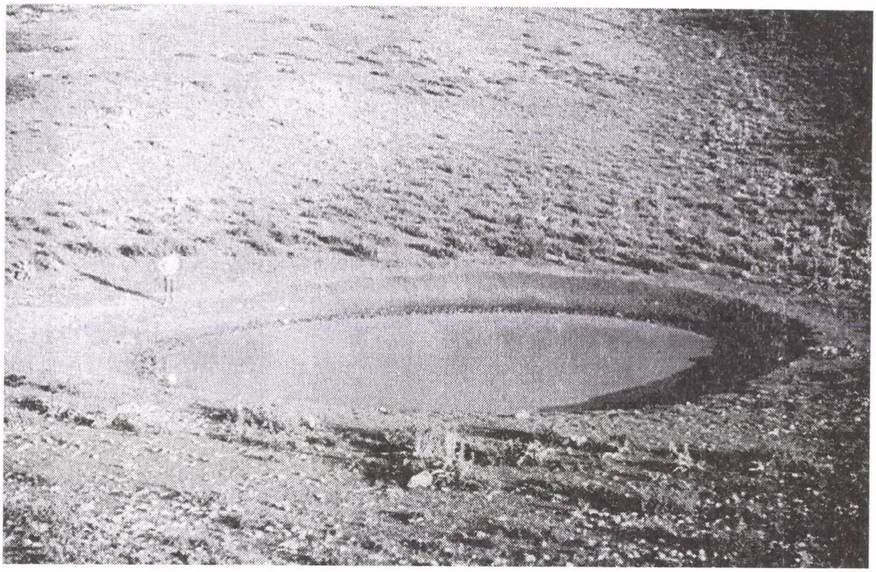

Photo 7. A „choked" doline, on the bottom of which the clayey sediments washed in from the sides have become compressed into an impermeable layer that prevents infiltration of the precipitation water in the centre of the doline. In this stage, such „lake" dolines no longer deepen, but only widen laterally

from the intermediate ridges between the dolines, which also play the role of local sediment-catchment basins. The relative deepening of the doline is accelerated further by the circumstance that, simultaneously with the washing-away of the soil from the ridges and saddles separating the dolines, these ridges become increasingly more prominent, as the dynamism of karstification is slowed down there in parallel with the almost automatic bleaking processes.

It should be noted that the *cessation* of deepening of dolines may also be caused by the otherwise essential washing-in of the soil. If a large amount of soil accumulates at the bottom of a doline, it may become compressed into *an impermeable layer* which prevents the further infiltration of precipitation water into the depths. In such a case the precipitation water no more comes into contact with the limestone through the soil deposited on the bottom of the doline; instead, it rather does so along the rims of the doline, where the soil thins out. The zone of intense dissolution in the old doline is therefore restricted to a ring-like area around the edge, and this results almost exclusively in the growth of the doline in the lateral direction, i.e. in its widening. One of the most illustrative examples of such a „choked" doline in Hungary is the Vörös-tó (Red Lake) on the Aggtelek Karst *(Photo 7)*.

Naturally, in certain karst areas there do exist „collapse" dolines over cave-in cavities too, such as the famous Macocha on the Moravian Karst or the huge gorge of the Skocijan cave in Slovenia. However, with their elliptic steep rock walls these abiogenic forms can always be distinguished precisely from the bowl-like cauldrons of corrosion dolines *(Photo 8)*.

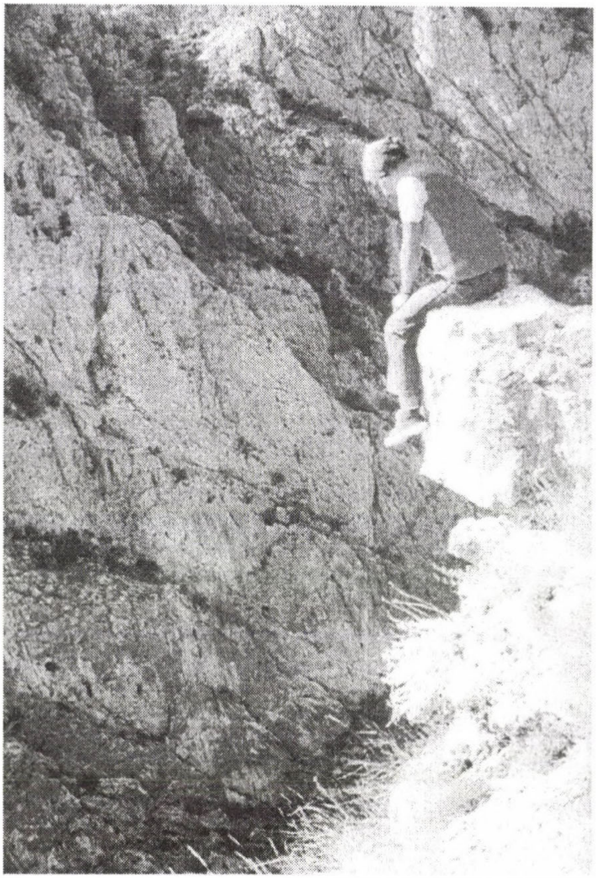

Photo 8. A characteristic steep-walled gorge doline near Imotski, Hrvatska. The several hundred metres deep, vertical walled rock cauldron was produced not by dissolution, but by the splitting up to the surface of a ceilingy of a deeply extending large cavern

Biogenic accumulative karst phenomena

Among the karst phenomena displaying biological regulation, however, not only dissolution forms are encountered. It must be recognised that the majority *of karst accumulation phenomenon groups* acquired the impulse for their formation, the dynamics of this process, and even at times the nature of the forms, from the activity of the biota. In caves, it is possible to find limestone depositions with the most varied appearances: calcite *stalactites* and *stalagmites (Photo 9)*, the various *encrusted dripstones; calcareous tufa (travertine) tetaratas* developing as steps in the cave streambed *(Photo 10)*; *calcareous tufa accumulations* occurring in surface valley

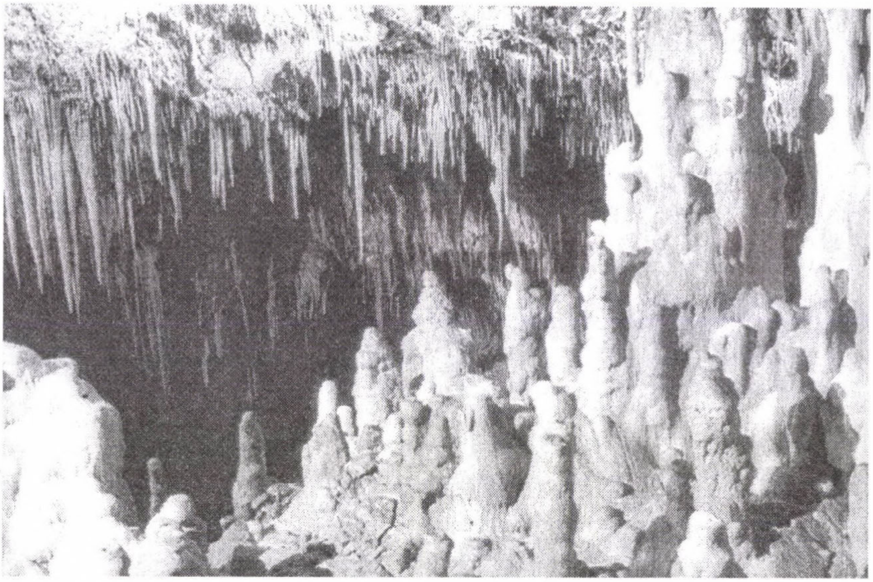

Photo 9. The dripstones of caves, stalactites and stalagmites, are similarly biogenic karst formations, as they are produced only in caves above which there is also biogenic karst corrosion on the surface. Calcite dripstones are unknown in the caves of barren, vegetation- and soil-free karsts of high mountains and in the frigid zones

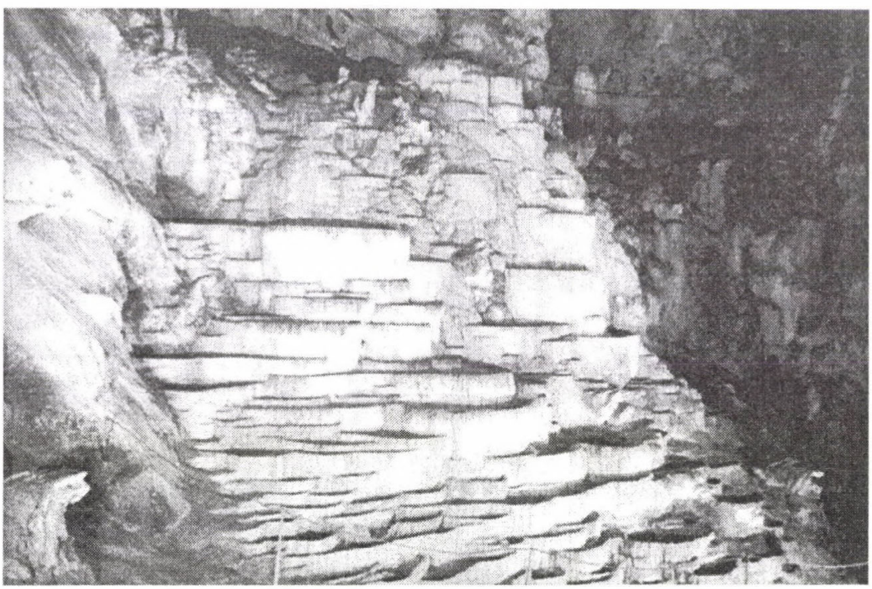

Photo 10. The calcareous tufa veirs and limestone tetaratas visible in caves are similarly products of the deposition of mineral substances dissolved up by biogenic corrosion. The carbon dioxide absorbed in the surface soil layer is lost from karst water infiltrating or trickling to the bottom of the cave, and consequently the limestone is redeposited in solid form

Photo 11. The surface calcareous tufa accumulating in the beds and waterfalls of karst streams are karst sediments of biogenic origin for two reasons: in addition to the dissolution saturation, the new deposition of the limestone is also accelerated by the vegetation. The carbon dioxide is extracted from the limestone solution by assimilation, by the green plants living in the water or swept into it

sections in the environment of karst springs, such as the travertine hill with its spring cave under the Palota-szálló (Palace Hotel) at Lillafüred in the Bükk Hills in Hungary, or the famous and beautiful waterfall tufa steps on the Plitvice Lakes in Croatia *(Photo 11)*; and even the *tufa curtains* on the hillsides of tropical karsts. In fact, these are all biogenic karst phenomena.

Nothing is changed in this classification by the fact that there are among them karst sediments displaying *indirect* biological regulation, where only the rock dissolution and the solution saturation aspects have been functions of the biological processes (e.g. limestone deposition in caves). They also include formations which reflect the activity of the living word both *directly* and secondarily, where the *way of precipitation of travertine* from solution is also regulated by plant assimilation, (e.g. the calcareous tufa accumulations of karst springs and karst streams).

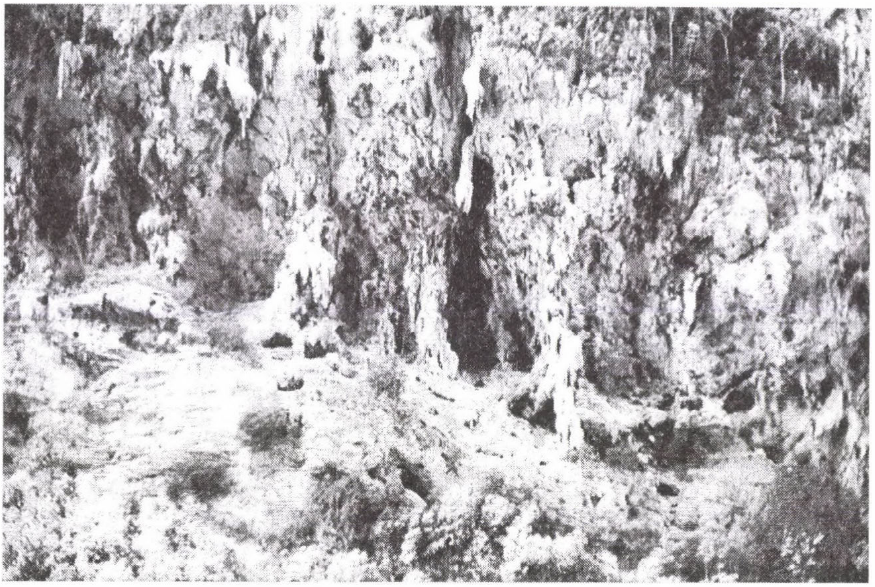

Photo 12. The tufa curtains of tropical karsts are formed by the soil moisture, with its rich calcium content, trickling under the vegetation growing on the steep hillsides, thereby encrusting and calcifying the vegetation. This is fatal for the plants, but the more intense the assimilation of the vegetation and the stronger its will to live, the faster and the more „relentless" will be the „suicide-like" accumulation of calcium

It is for this reason that there are no dripstones in the caves of the abiogenic-surfaced polar regions and of the high mountains free of vegetation, and *this is why* the karst springs and streams originating here do not deposit calcareous tufa in their beds. In tropical karsts with abundant vegetation, on the other hand, a wonderful variety of dripstone formations is produced even on the surface; these enclose and „petrify" the intermeshed green network of lianas, tendrils and the like clinging to the steep rock walls *(Photo 12)*.

To summarize what has been said, the genetic taxonomy of the most important karst phenomena have been tabulated *(Table 1)*.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 026493.

Table 1. Genetic system of the most important limestone karsts phenomena (L. JAKUCS)

Group I		Group II		Group III	Group IV
Column A	*Column B*	*Column A*	*Column B*		
Biogenic corrosional karst phenomena	Accumulation karst phenomena dependent on biogenic corrosion	Abiogenic corrosional karst phenomena	Accumulation karst phenomena related to abiogenic corrosion	Erosional karst phenomena	Karst phenomena due to tectonic movements and collapse
1.Lapies fields formed under the soil	1. Cave dripstones	1. Gravitational (precipitation or rainwater) lapies	1. Cave clay sediments	1. Karst valleys, limestone gorges, karst canyons	1. Rock fissures, fissure caves
2. Root lapies, channels	2. Cave tetaratas	2. Cave lapies	2. Hot-water springs limestone	2. Stream and river terraces	2. Collapse dolines
Column A	*Column B*	*Column A*	*Column B*		
3. Subsurface dissolution rock cracks and small cavities	3. Cold-water spring limestones	3. Vertically-developed small caves of higher karst plateaus	3. Hydrothermal karst minerals (gypsum, aragonite, barite, etc.)	3. Most poljes	3. Cave halls formed by collapse
4. Certain avens	4. Surface tetaratas, calcareous tufa formations	4. Hydrothermal caves		4. Natural arches	4. Certain avens
5. Dolines, doline series	5. Surface drip-stone formation, tropical tufa curtains	5. Hydrothermal effects on rocks (crumbling)		5. Ponors (blind-valley swallow holes)	5. A smaller part of poljes
6. Uvalas				6. Active and inactive stream caves	
7. Mogote				7. Multi-level cave systems	
				8. Flow depressions, evorsion cauldrons, cave terraces	
				9. Cave sand, gravel and boulder sediments	

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ECOLOGICAL INVESTIGATION ON SOME HUNGARIAN KARSTS

ILONA -BÁRÁNY-KEVEI[1]

Introduction

The sensitivity of karst systems to human activities has become increasingly apparent as a result of research during the 1980's and 1990's. The environmental impacts on karst regions must be analysed, since these processes take place very rapidly. Non-karstic materials integrate quickly into the karst water system, modifying or damaging the natural forms that have been developing for millions of years. Karsts are therefore especially sensitive geo-ecological systems and research on different aspects has been encouraged since the 1980's (JAKUCS 1980, 1987; BÁRÁNY-KEVEI 1976, 1985a,b, 1987; PFEFFER, 1990; HARDWICK and GUNN 1996; TRANTER, GUNN, HUNTER and PERKINS 1997).

The paper presents some results of research into karst-ecological systems in some Hungarian karst areas.

Methods

The general research methodology adopted is applicable to all kinds of karst regions. When investigating the factors of the system (soil, microclimate, vegetation and microbial activity) the methods of the scientific fields can be applied respectively.

The parameters of the soil samples from the outcrops were analysed in laboratory: grain composition (aerometrical analysis), carbonate content (Scheibler's calcium-meter), pH value (digital pH meter), hydrolithic acidity (titration), heavy metal content (Perkin-Elmer atomic adsorption spectrophotometer). Nutrient analysis and the definition of the water soluble ions were carried out at the MÉM NAK Institute at Hódmezővásárhely according to Hungarian standards.

Microclimate monitoring was also undertaken at each site. The following parameters were measured: soil temperature at 5 and 30 cm (electric resistance thermometers), sunshine hours (Campbell-Stokes radiation meters), wind velocity (anemometer), air temperature (Assmann's psychrometers).

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In each doline vegetation was surveyed using one square on the northern, eastern, southern and western sides of the slopes each, and one on the bottom. Both the plant species and percentage cover were recorded. The karst vegetation was evaluated using the ecological indicators (water budget, heat budget, soil reaction and nitrogen demand) according to B. ZÓLYOMI (1966) based on the known community composition. A survey of the microflora (defining the number of aerobic and anaerobic bacteria on Agar nutritive soil) was carried out at the Microbiological Department of the University of Szeged.

Discussion

Karst dolines provide excellent basis for the study of these interrelated processes. The first sphere of the karst-ecological system is the air just above the surface, where there is a karst microclimate. While macroclimate is responsible for the quantity and intensity of precipitation, microclimate modifies the quantity of water infiltrating into the rocks. The latter also affects the development of vegetation, influencing the quantity of $C0_2$ produced during root respiration and regulates soil temperature and humidity. Millions of micro-organisms living in the soil change the components of the soil atmosphere through the decomposition of organic materials and through their own metabolism. They also make an indirect effect upon the physical and chemical properties of soils and exert an influence on the quality of seepage water, leading to corrosion processes of different scales. The inner dynamism of soil can prevent extreme changes to occur in the system (through buffer ability and redox potential), however, it cannot compensate for the processes of long duration. The inner dynamism of soil can change on the long run, leading to a possible malfunction of the whole system. The changes provoked by external forces are reversible down to the contact with the bedrock. However, as soon as waters enter the parent rock, they become irreversible. Water in the rock layer (the aquifer) is the primary transporting agent of matter and energy. This water reaches the earth surface again in the form of karst springs and if polluted, its value for economic purposes and for the maintainance of the ecological balance is decreased. Another irreversible process, the degradation of dripstone might also occur owing to polluted water entering the cave.

The survey of karst dolines is especially important, since these depressions are the most endangered places in many karst systems because they provide concentrated recharge.

The present study compares some areas in Hungary: Aggtelek Hills, Bükk- and Mecsek Mountains. Most of the Mecsek and Villányi karst areas have been affected by human activities (coal mining and quarrying). In contrast, the karst region of the Bükk Mountains almost completely retained its original character, owing to the protective measures within the Bükk National Park. This has been a nature conservation area for a much longer time (23 years) than any other karst region. The Aggtelek National Park exists for 13 years, thus human impact is relatively minor. However, as a result of grazing, other farming and forestry activities prior to the conservation act, some traces of human impact are still present. Now that Baradla Cave with its Domica Branch on the territory of Slovakia have become part of World Heritage (1996), geoecological investigations became necessary in the framework of landscape conservation.

The *relationship between soil, microclimate and vegetation* was investigated, because these components influence the entire karst system.

Physical and chemical characteristics of karst soils

These features are of importance from the aspects of the stability of karst ecosystem. Soils are capable to buffer extreme impacts, in the case of drastic effects, however, they can even intensify them, since the enzymes getting into the soil stay there for a long duration. Soil dynamism is primarily expressed in its chemistry which influences the development of soil aggregates (structural soil elements). The structure and texture of soil define its air, water and heat budgets. However, chemical and physical features are bound to change in relation to biological activity.

In Hungary the *physical quality* of karst soils is poorly differentiated, they are predominantly unconsolidated, immature, having been primarily developed on solution residue or on loess-like sediments. Their dominant fraction is loam (50–60%), while the sand fraction is poorly represented in the Bükk dolines. The soils on the Aggtelek Karst are poorly sorted, too. Their clay content is 20%, higher than that of the soils in the Mecsek karst region. This considerable clay content is due to the older dolines at Aggtelek than those in the Mecsek Mountains. These soils have considerable water storing capacity. The thick clay rich sediments can eventually become impermeable and in the karst depressions this clay sedimentation makes the karst corrosion effects move towards the rims so the dolines become wider rather than deeper.

The pH, hydrolythic acidity, alkalinity and the $CaC0_3$ content describe the *chemical state of soils*. The water soluble anions and cations, being important in karst corrosion, also represent the chemical properties of soils.

The soils in the Mecsek and Aggtelek Mountains are slightly acidic (average pH 6.0–6.5). Of them the soils at Aggtelek are more acidic (0.3–0.4) (BÁRÁNY-KEVEI, 1992). At several sites values of 5.0 were recorded indicating the acidification process under way (in the summer of 1995 *Calluna vulgaris* was found on the karst surface at Aggtelek). At the same time it is known that soils formed on limestone are generally non-acidic. To investigate occasionally low pH values the difference between the pH of soil solutions with water and with potassium chloride were measured. An increase in the difference between the two pH values indicates acidification of the soil as it has been the case in the dolines of the Bükk, Aggtelek and Mecsek mountains alike. Since there has been a trend for acidification recorded at sites where direct human impact is low (e.g. in the Bükk National Park), it can probably be regarded the result of acid deposition.

Analysis of *water soluble cations and anions* can be used to describe the chemical state of the soil. Naturally, the Ca^{2+} ion content is high in karst soils, but the K^{+}, Na^{+} and Mg^{2+} ions are also abundant. Anions SO^{4--} are abundant but there are many

Cl^- too (BÁRÁNY-KEVEI, 1987). There is a general tendency for both cations and anions to be found on the slopes near the rims of the dolines, rather than at their bottoms. The minimum quantity of ions deposited at the bottom indicates an intense leaching process in this level.

The large volume of *heavy metals* presented in the soils may indicate ecological change. Heavy metal concentration was analysed on the karst area of Peak District in England by COLBOURN and THORTON (1978) and by XIANDONG and THORTON (1993). They suggest the enrichment of metals within the soil is affected by nature of the underlying mineralised rock, mining activities and pollution by smelters.

Heavy metal analyses were performed on soil samples from Aggtelek, Bükk and Mecsek mountains. Soil samples were collected from a single doline in each area.

The values are too high in all the three of karst samples so it is improbable that they originate from the rock alone. The limestone is relatively pure, the heavy metal content of karst soils comes from the rocks only partially. The source of contamination by metals are dry and wet deposition.

Intensity of microbial activity

Microbial activity is affected by the pH and ion content of the soil. There are millions of soil microbes (bacteria, ray fungi and microbial fungi) in 1g of soil. More than two thirds of the $C0_2$ emitted in soil is a result of decomposition of organic material, and only one third comes from root respiration. The most intense $C0_2$ production takes place in the upper 20–30 cm soil layer, as a result of the aerobic bacteria activity. This is dependent on the temperature and humidity regime of the soil. If temperature is low (e.g. lower than 20 °C) bacterial activity decreases. Low soil humidity (e.g. below 20–30 volume %) also reduces microbial activity. The dynamism of the soil is thus a very complicated system and the least change in any of the factors may result in change due to positive feedback in whole of the karst system.

The soils from Bükk and Aggtelek (in both mountains 16–20 samples from each doline were analysed) were not found to favour microbial activity (KEVEI-BÁRÁNY and ZÁMBÓ, 1988). In soil under forest (brown earth like soil) there is typically 15–20 million bacteria in 1 g, being much less in grassy ecosystems (in rendzina like soil). There is a significant relationship between the number of bacteria and soil temperature in the surface layers of the doline soils. In deeper layers, however, the number of bacteria correlates with humidity rather than temperature.

Researches performed in the summer 1995 in the Aggtelek Karst demonstrated that the number of microbes near the surface of the soil than the deeper, and the microbial activity is more intensive in forest association than elsewhere. The quantity of $C0_2$ emitted by microbes is a real ecological entity. Three soil profiles levels were investigated (1/1 = near-surface, 1/2 = middle level of the profile, 1/3 = lower level of the profile) are situated in the upper 60 – 70 cm of the soil.

Impact of the vegetation

Vegetation cover exerts a strong influence on the processes in the soils of the karst. The karst shrub woods (Orno-Cotinion) are characteristic of Hungarian karst areas. The alteration of the Central European mountainous beech wood (Fagion medio-europaeum) covers the karst surface above 700 m elevation in the Bükk Mountains. Its vegetation typical of the Hungarian middle mountains stretches down into the oak belt. There was a extensive clearing of the forests in the Hungarian Mountains early in this century. Barren karst surfaces, which appeared after deforestation can still be recognised in some areas, but they are not common in Hungary. The only traces of deforestation can be seen in the very slow natural regeneration of forests of the dolines. In most of the dry valleys juniper appeared following deforestation and demonstrates the low nutrient availability within the soil. The diverse doline vegetation is increasingly becoming more uniform. Grazing has significantly contributed to the decline of biodiversity.

The species composition of the vegetation in the Bükk dolines reflects the common features of the karst. The species mapped included those characteristic of the mountainous and submontane beech wood as well as steppe meadow, rocky and puszta grassland slopes, tufted grass and montane hayfields. The average values of the ecological indicators (water and heat budgets, soil reaction and nitrogen demand) are presented on the basis of sampling dolines in the Bükk Mountains. There is no significant differences between the heat budget and the vegetation species found in the Bükk dolines, but since this area is a micro region, even the 0.45 difference cannot be neglected. The higher heat budget value recorded on the northern slopes demonstrates the aridity of these slopes and influences all other ecological factors.

The effect of the exposure is also clear in the analysis of the water budget indicator. Its average value is 6.62 in the northern, and 2.82 in the southern parts of the dolines, showing the slope-dependent distribution of soil humidity. Like heat and water budgets, the differences in soil reaction are also significant (BÁRÁNY-KEVEI, 1985).

The ecological values measured at Aggtelek differ from those measured in the Bükk Mountains; the species requiring less nitrogen are more abundant at Aggtelek. It is due to the intensive grazing, having increased the nitrogen content of the soil in the former region. Many species present were not members of the original plant community. Average indicators of the water budget show temperate-fresh and temperate-dry characters.

The effect of microclimate

The dominant factor of karst formation and development is *climate*, but the ecological factors are influenced to a greater extent by *microclimate*. In the Hungarian karsts microclimatic systems exert their influence through specific orographic and morphologic condition. The independent microclimatic areas of the karst dolines are the most characteristic where the microclimate modifying effect of the exposure prevails side by side with the effect coming from the enclosure of the depression (BÁRÁNY, 1976, BÁRÁNY-KEVEI, 1985). The differentiated warming up of the different slopes according to aspect results in important differences in the energy input and temperature of the soil. Differences in temperature affect both the microbial activity and the composition of the macroflora. Temperature conditions of the western and north-western slopes are found to meet the demands of bacterial activity the best. The desiccation of soils prevent the bacterial population from significant increase, due to the high humidity and low temperature of the northern slope, and the strong radiation input of the southern slope. As far as the whole ecological system is considered, this results in a slower decomposition and transport of organic materials on the southern slope than on the other ones.

Conclusions

The karst-ecological system may be defined as a structured and dynamic system in which rock, soil, microclimate and macroclimate represent the abiogenic elements, while macroflora and macrofauna, microflora and microfauna represent the biogenic ones. The interrelationship of the biogenic and abiogenic elements, along with the material and energy flux occurring in this interrelationship maintain the evolution of the system. Its structure is defined by the vertical and horizontal distribution of its elements. Its specific features include its sensitivity, the velocity of its processes and its three-dimensional surface of effects.

1. The physical and chemical characteristics of karst soils are important the change of karst ecosystem. Soils can buffer extreme impacts, though in case of very strong influences they themselves serve as agents in intensifying impacts, may be either advantageous or deleterious. Since the enzymes of micro-organisms get into the soil and remain there for a long time, the inner dynamism of soil is independent.

The difference between the two pH values (in soil solution with water and potassium chloride) exceeds the limit indicating acidification in the dolines of the Bükk, Aggtelek and Mecsek mountains alike. The higher the clay and organic material content, the more heavy metals are bound on the colloids. Neutral chemical reaction also supports the absorption of heavy metals. In strongly acidic soils most of the metals enter solutions.

2. Biological processes have a feedback on the chemical properties of the soil through the decomposition of the humus materials, so the maintainance of the natural bacterial populations and conditions is desirable. There is a close relationship between the number of bacteria and soil temperature in the subsurface layers of the doline soils. In deeper layers, however, the number of bacteria correlates with humidity rather than temperature.

3. The vegetation developed on karstic rocks with rendzina and clayey soils of forest. The species develop associations here that can adopt themselves to the extreme water budget of the soil. If the vegetation changes, as at Aggtelek, both the intensity of karst corrosion and the further functioning of the karst ecosystem are subject to change. The degradation of vegetation is acting against natural processes, as shown in the appearance of a few heather species along the edges of the dolines. Their extension is limited, but they are the environmental indicators of the change.

4. Microclimatic systems which modify the radiation impact are formed within the local mountainous and valley climates under specific orographic and morphologic conditions in the karsts. The differences in daytime radiation input depending on slope aspect are not compensated for by the night-time heat emission, since the flow of the cold air causes cooler air to be accumulated in the dolines (thermal inversion). This microclimatic feature results in the specific inverse distribution of vegetation which is lower in the bottom of the doline than along its edges.

The future exploitation and management of the karst areas has to rely on the knowledge about the function of the karst-ecological systems. This knowledge can only be acquired through methodologies developed by landscape ecology.

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DEMONSTRATION OF PAST SEDIMENT COVER ON PRESENT-DAY KARST LANDSCAPES

MÁRTON VERESS[1]

Abstract

Boundaries of former covered and uncovered karsts can be determined through the analysis of the features developed along rock boundaries since the latter survive the denudation of the sediment cover. Not only the conditions of coverage in a karst area i.e. the previous existence of uncovered areas but also their boundaries can be determined through the detection of rock boundaries.

Since the altitudes of rock boundaries framing uncovered karsts were nearly identical, the different elevations of the neighbouring uncovered areas and the extent and nature of the uplift of a single uncovered karst area could be deduced from the present altitude differences of the features developed along rock boundaries.

Introduction

The hydrological relationship between the karst and its surroundings is paramount in the evolution of karsts.

L. JAKUCS (1968, 1971) distinguishes between authigenic and allogenic karsts. In the former case waters flow out of the karst area, whereas in the latter case waters flow into the karst. Stream valleys on allogenic karsts can continue on the limestone (JENNINGS, I.N. 1971).

Valleys are most frequently missing on authigenic karsts due to infiltration and lack of weathering products (CHOLNOKY, J. 1944). Even if they develop, valley formation might be triggered by:

– Collapses of caving-in within the karst (CHOLNOKY, J. 1944).

– Climatic conditions during glaciations in periglacial areas when waters cannot infiltrate into the frozen soil (C. REID, 1887).

– Solution process might also be involved in the valley formation (WILLIAMS, P.W. 1983).

– Through the fusion of karstic forms.

– Through erosion, by the emerging springs (SPARKS, B.W. and LEWIS, W.V. 1957).

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– Having developed in impermeable sediments covering the karst, the valley then cuts through the latter and proceeds in the limestone until the sediment cover is totally denuded (LÁNG, S. 1955, WARWICK, G.T. 1964, DÉNES, Gy. 1971, JAKUCS, L. 1971, HEVESI A. 1978). Valley development may continue after the complete removal of the sediment cover if the karstic water table lies close to the channel floor and the incision of the channel follows the sinking of the water-table (HEVESI, A. 1984, 1986).

Epigenetic valleys only occur in karst areas formerly covered with sediments. Considering the epigenetic (superimposed) valleys of Bükk Plateau (Hungary) A. HEVESI (1980) recognised the drainage network having developed in the covered karst stage of the plateau.

Taking into account that the epigenetic valleys of a karst region indicate the extension of the former sediment cover the margins of this cover can be detected (VERESS, M. 1993) in the case if the uncovered limestone relief has higher altitude than the covered had. The boundaries of the former sediment cover can be also detected where the covered relief had the higher altitude since karstic features develop along such rock boundaries.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 024162.

Detection of margins of the sediment cover

Some parts of karsts become covered when due to former peneplanation and tectonic dissection (or both) surfaces develop at different altitudes. The emergent isolated authigenic karst sections are surrounded by rock boundaries of different type (see below).

If the latter are detectable after the denudation of the sediment cover, not only the former uncovered karstic topography can be detected but the boundaries of the latter as well. Since the sediment cover developed at similar altitudes the latter appeared along rock boundaries referring to the subsequent uplift of the area. The alteration of altitudes within short distances of former rock boundaries on uncovered reliefs indicate the development of a fault structure, while alteration within longer distances refers to tilted uplift. Rock boundaries of different altitudes of the neighbouring uncovered areas indicate different uplift of the areas.

Rock boundaries can be active and inactive (VERESS, M. 1990, VERESS, M. and FUTÓ, J. 1991). If the covered relief slopes towards the uncovered one the rock boundary is active. Sink holes are the characteristic formations of such rock boundaries. Following the denudation of the sediment cover swallow holes (DÉNES, Gy. 1971, JAKUCS, L. 1971, HEVESI, A. 1978, 1980, 1986), doline ponds (JAKUCS, L. 1971) avens (DÉNES, Gy. 1971, HEVESI, A. 1984), and wallowing places (VERESS, M. 1991) can develop from these sink holes. If the water table is close to the uncovered limestone relief water flows are not captured at the active rock boundary and sink holes do not develop. Therefore valley formation takes place on the uncovered relief as well. In this case the detection of the former rock boundary is not possible unless the water flow is captured in time after all. (It is probable if the sinking of the water table is faster than the deepening of the valley floor.) In this case, due to the retreat of the rock boundary, the detectable sediment cover is smaller than its original extension.

If the relief slopes from the uncovered karst towards the covered one the rock boundary is inactive. This results in epigenetic valley formation in the lower situated covered karst areas. The heads of the superimposed valleys in the limestone retreat toward the inactive rock boundary. Such former rock boundaries are marked by the heads of the epigenetic valleys *(Fig. 1)*.

In both cases valley development might proceed on the uncovered relief as well even if the water table lies on the karst surface (or close to it). However, in karst regions where the uncovered parts show isolated protrusions in the sediment cover water table lies deep below the surface of the uncovered relief. It means that the altitude of the sediment cover determines the altitude of the water table. Thus valley development cannot occur on uncovered areas because water flows are captured. In the described situation water table lies close to the surface only if perched water table develops above the regional water table. It is possible if the carbonate rocks alternate with non-karstifying impermeable rocks. Therefore the presence and the boundaries of uncovered karst protruding from the sediment cover can be detected if the adjacent swallow holes, avens, wallowing places (former active rock boundary) and the valley heads of the adjacent epigenetic valleys (inactive rock boundary) are connected with a curve on the map. However, it may happen that the former uncovered karst cannot be confined from all sides. That is the case when the direction of the inclination of the former covered relief and that of the strike of the rock boundary coincide, epigenetic valleys develop parallel to the rock boundary *(Fig. 2)*. Then the former inactive rock boundary blocks the development of valley heads on the carbonate relief. This rock boundary section, around the uncovered karst can be approximately detected with the help of the altitudes of the rock boundary sections already reconstructed, since the level of the cover has a similar altitude. (Naturally, it has to be taken into consideration that for tectonic reasons some rock boundary sections could be located at different altitudes.

Swallow holes, avens and wallowing-places can form a row or rows inside the previously uncovered karst in the direction of the inclination of the surface (on subhorizontal areas, peneplains).

This morphological situation suggests a sediment cover to have formed in a narrow zone inside the uncovered karst. In such case the epigenetic valleys will have the same direction as the strike of the sediment cover in a narrow zone. As a result, the rock boundary develops inside the valley. The retreat of the rock boundary in the valley results in linear arrangement of the karst features *(Fig. 3)*.

Description of the model area

The following model areas have been chosen in the Bakony Mountains (part of the Transdanubian Mountains, Hungary): Som Hill, Papod-Borzás and Hajag *(Fig. 4)*.

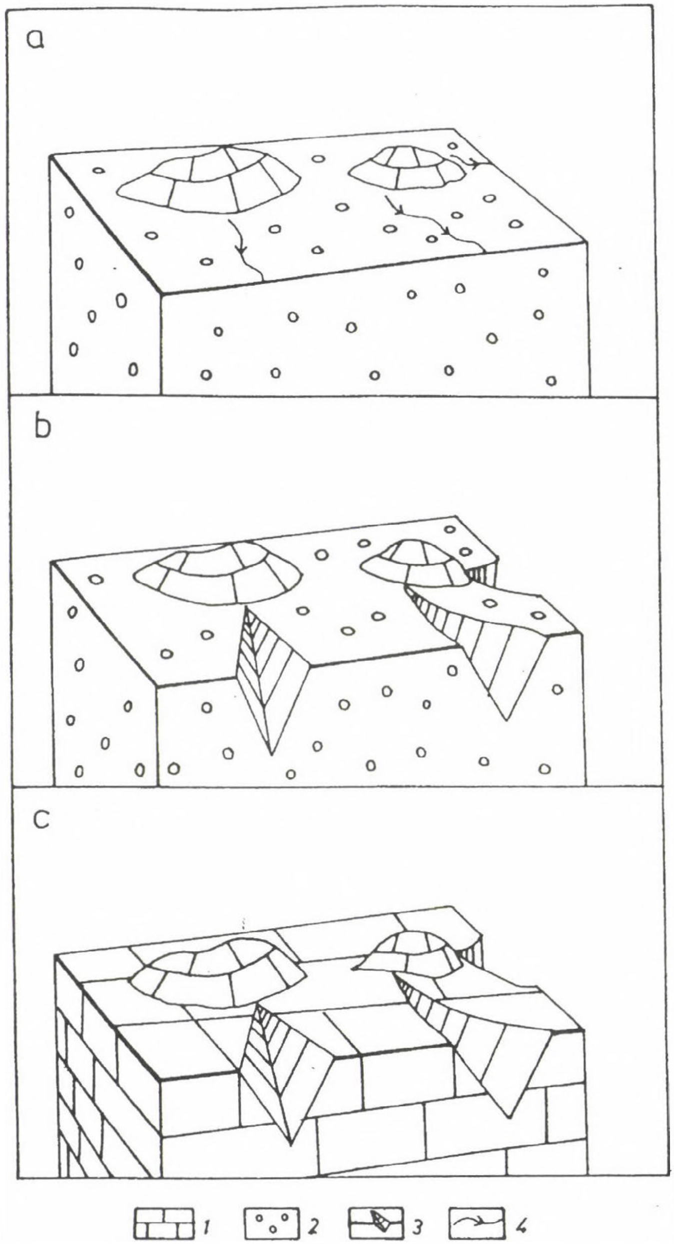

Fig. 1. Development of valley heads at inactive rock boundaries. – a) limestone relief partly covered by non-karstic sediment; b) valley formation following the elevation; c) karstic relief with epigenetic valleys and lost sediment cover. 1 = limestone; 2 = sediment cover; 3 = valley; 4 = water-course

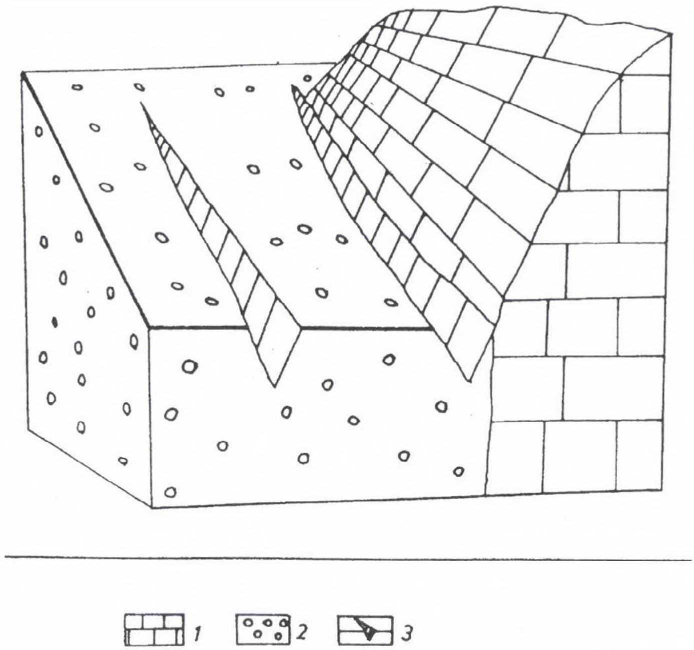

Fig. 2. Inactive rock boundary without valley head formation. 1 = limestone, 2 = sediment cover; 3 = valley

Peneplain formation took place in the Bakony Mountains from the Triassic until the end of the Cretaceous (tropical karstic peneplain), then since the Eocene it has been dismembered as a result of the tectonic movements (PÉCSI, M. 1980, 1987). The horst and graben terrain developed in this way was subsequently inundated during the Middle Eocene marine transgression (PÉCSI, M. 1987). Due to the dissection of the peneplain this first burial was not uniform (the developed sediment cover consisted mostly of limestone and less marl). The second burial of the peneplain lasted from the Oligocene until the Middle Miocene (KORPÁS, L. 1981). Then, during fluvial sediment formation (impermeable) clastic sediment cover developed. The particle size in Csatka Gravel Formation varies from gravel to clay fraction. Various parts of the peneplain could have been situated at different altitudes since the diameters of the fragments of the sediment mass of Csatka Gravel Formation show spatial differentiation. During the Quaternary uplift of the mountain (RÓNAI, A. 1983) Csatka Gravel Formation was strongly denuded, and at present it can be traced only in patches. Parts of the tropical karstic peneplain had dismembered into blocks showing a rather different evolution during the Cainozoic (PÉCSI, M. 1980, 1987).

The blocks were classified by M. PÉCSI (1987) on the basis of their former and present geomorphological position. Mountains being in elevated position during the Cainozoic (and at present as well) are called horsts. They are composed of Mesosoic (overwhelmingly Upper Triassic) carbonate rocks. So the study of the sediment

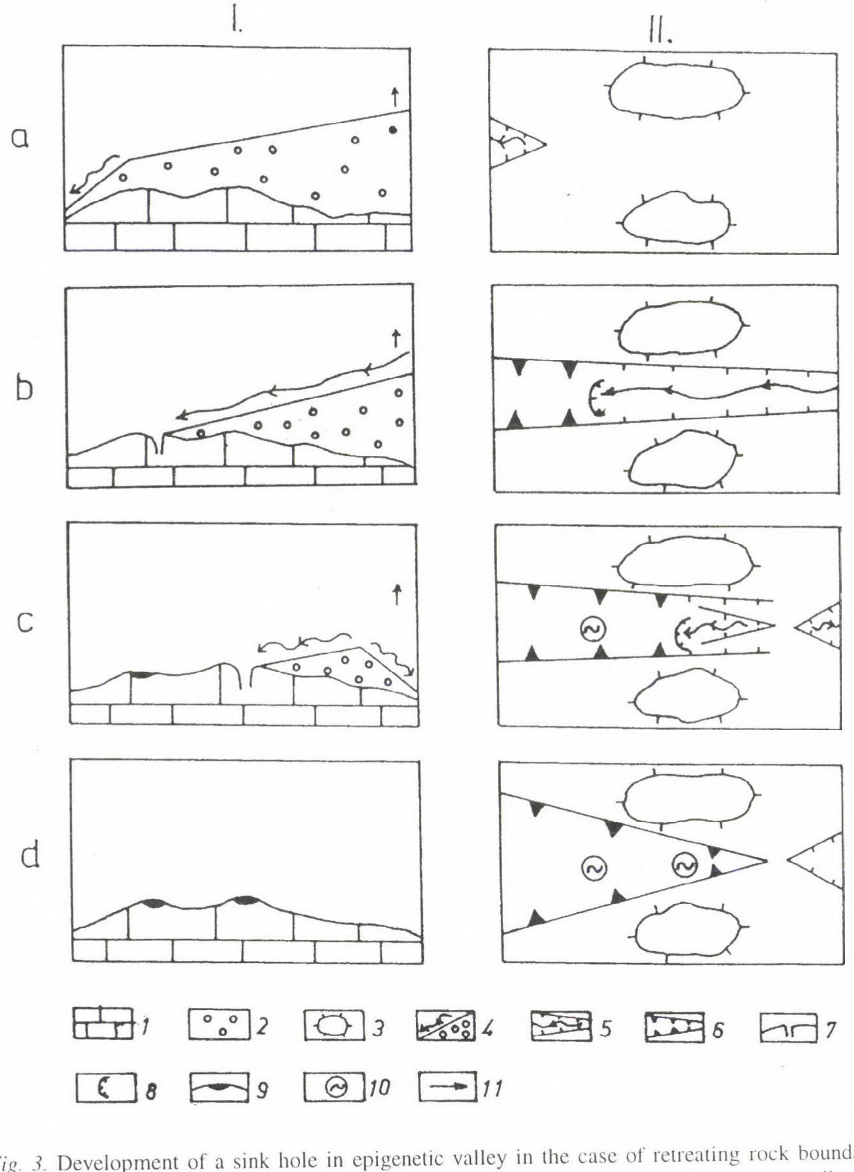

Fig. 3. Development of a sink hole in epigenetic valley in the case of retreating rock boundary. – I/ side-view; II/ top-view. a) Valley formation on a covered mountain top declining in one direction; b) continuing valley formation; development of a sink hole; c) valley formation on covered mountain top declining in two directions, development of a subsequent sink hole; d) mountain with inactive sink holes and lost sediment cover. 1 = limestone; 2 = sediment cover; 3 = area protruding from the sediment cover; 4 = side-view of a valley developed in sediment cover; 5 = top-view of a valley developed in sediment cover; 6 = top-view of an epigenetic valley; 7 = sink hole from side-view; 8 = sink hole from top-view; 9 = inactive sink hole from side-view; 10 = inactive sink hole from top-view; 11 = elevation

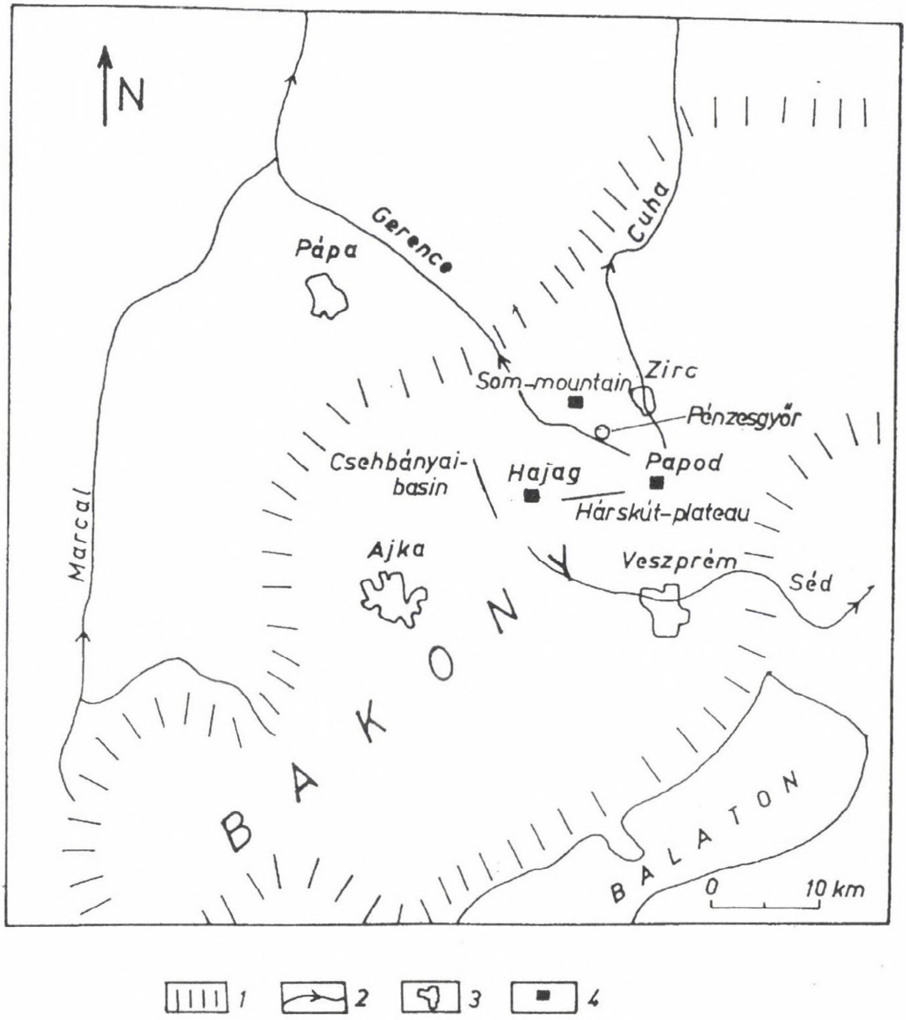

Fig. 4. Model areas in Bakony. – 1 = mountain margin; 2 = river, stream; 3 = settlement; 4 = investigated area

covers of these horsts has practical relevance. That is, marine transgression and sedimentation in the Bakony Mountains preceding the present partial non-karstic (loess) sediment cover is the least expected in the area of the horsts. Moreover, no perched water table could develop inside them (Upper Triassic carbonate rock sequences do not contain impermeable rocks).

The sediment cover in the Bakony Mountains on which the epigenetic valleys developed (LÁNG, S. 1958) was formed during the second burial. Therefore its examination in the area of the three horsts (Som Hill, Papod-Borzás, Hajag) helps understand the extension of Csatka Gravel Formation.

Covered karsts developed in the Holocene, partly on the surfaces mantled by loess (VERESS, M. 1982, 1985, 1991). Present forms of these karsts are not suitable for the detection of the former active rock boundary.

In these areas two types of valleys occur (PÉCSI, M. 1991). Shorter valleys were cut by intermittent streams flowing along structurally preformed linear surface features (e.g. fissures) or incised in loess also due to erosion. Some of them can be of corrosional origin. Several kilometers long, wide and meandering valleys overhang the hillsides of the horsts. Their lower parts proceed on the relief still consisting of Csatka Gravel Formation which proves their epigenetic evolution and their development being related to this sediment cover. Quaternary or Pliocene sediments occurring on several valley floors (CSÁSZÁR, G. *et al.* 1981) testify to the development of the valleys having already begun before the Quaternary uplift of the Bakony Mountains. It is also proven by their incised meandering caused by the gently sloping surface. These valleys developed along the meanders during the uplift.

For the detection of the inactive rock boundaries these two types of valleys are considered below.

Evaluation of the reconstructed maps of sediment cover distribution in the model areas

The case of probable former burial and exhumation

The margin of the former burial on Som Hill is indicated by the valley head at 620 m altitude starting out to the north and the valley heads at 590 and 580 m opening to the east. The Nagy Pénz-lik and the Kis Pénz-lik (caves) and the adjacent inactive sink holes among the surrounding hills on the summit level of the Bakony Mountains indicate that the burial had developed in this zone.

The sink hole and the doline on the northern side (west of the valley with its head at 620 m) make the border uncertain to the west. It is likely that on the western side of Som Hill, as well as at other places in the surroundings, the terrain formerly sloped northward. Thus the slope of the covered relief and the rock boundary might have similar orientation. The rock boundary here can be identified with the help of the altitudes of rock boundaries reconstructed on the other slopes of the mountain. The western boundary of the former burial of Kis-Som Hill is marked by the valley heads at 590 m and by two inactive sink holes (at the same altitude).

The valley heads are situated at 560 and 570 m altitudes in the north. There are some inactive sink holes (wallowing places) at the latter. In the east a valley head at 500 m and an inactive sink hole at 510 m are found. By the use of the above data a map representing the former burial of Som Hill can be drawn *(Fig. 5).* It is probable that three elevated surfaces emerged from the burial formerly, the two elevations of Som

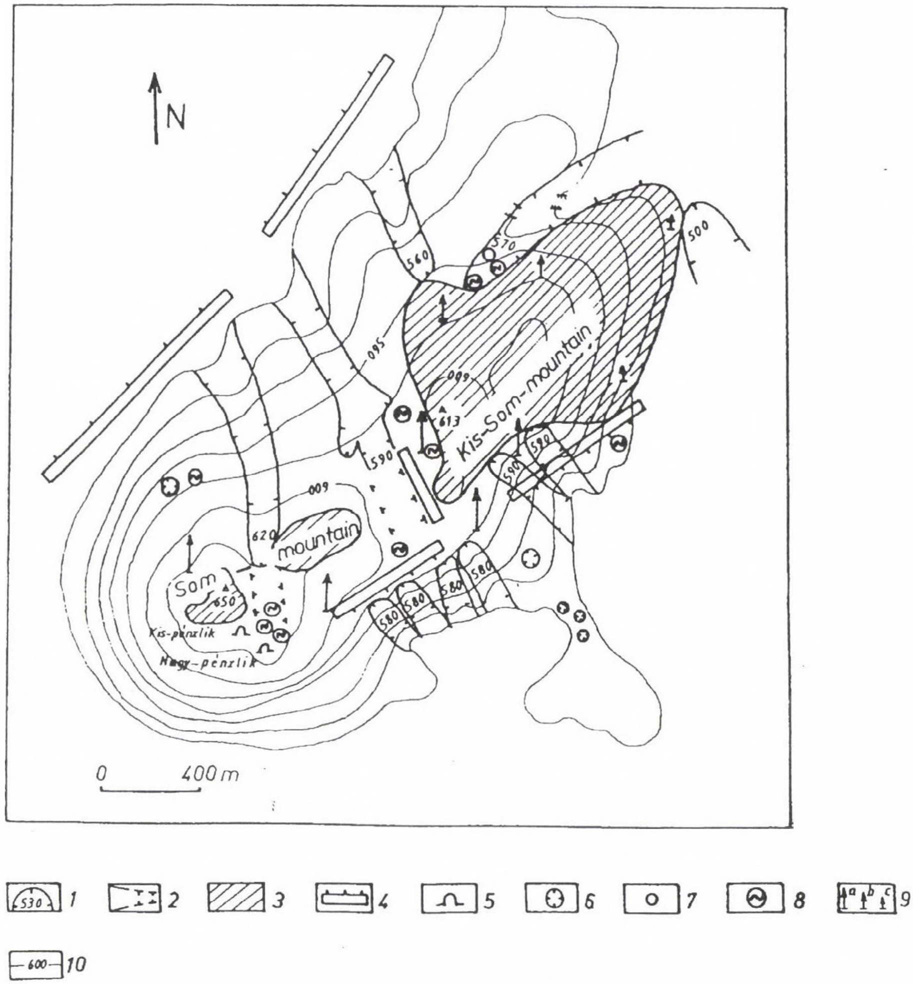

Fig. 5. Reconstructed map of the burial of Som Hill. – 1 = valley head; 2 = valley with its former upper part; 3 = former uncovered area; 4 = fault; 5 = aven; 6 = doline (solution doline, subsidence doline); 7 = inactive sink hole; 8 = wallowing-place; 9 = degree of elevation: a) high; b) medium; c) low; 10 = contours

Hill and the higher part of Kis-Som Hill. The saddles between these elevations were buried.

Rock boundaries occurred in the lower northern parts of the valleys having developed in these sediments as the streams incised. The first sink holes developed along rock boundaries.

Later these sink holes had turned into caves such as Nagy-Pénz-lik. The latter developed on the southern slope of Som Hill which means that the catchment of these caverns must have been located in the south of Som Hill. The former covered relief sloped from south to north since formations (active and inactive) indicating the former rock boundary are absent on the western side of the Bakony and the former catchments of the north-south oriented valleys developed at lower altitudes were found in the southern part of the mountains.

Due to the partial denudation of the sediment cover on the mountains and in their surroundings, probably owing to the deepening of the Gerence stream, the surface sloped not only northward but southward as well. Valleys on the southern side of Som Hill could also develop. These younger valleys are significantly smaller (shorter and narrower) than the valleys on the northern side. On the western side of Borzás the heads of valleys stretching out to north and south from the saddle are located at 570 m *(Fig. 6)*. On the southern side of Borzás three valleys begin at 470 m indicating the former burial.

The valley between Borzás and Papod and its side valley ascends up to 510 m. Two other valley heads of the southern side are to be found at 480 m. On the northern side under the Borzás there are valleys stretching up to 540 m.

The inactive karstic features in the saddle between Borzás ad Papod indicate the former burial. Valleys stretch up to 580 m on the northern side of Papod (the easternmost valley extends up to 560 m).

Using the above described configuration the distribution of the burial on Papod-Borzás is shown in *Fig. 6*.

With the exception of the saddle between them, Papod and Borzás probably elevated from the burial. A row of inactive karstic depressions refers to a valley which could develop on a relief with sediment cover. South of Papod-Borzás the surface of the saddle was entirely buried by sediments sloping northward. During further development (similar to that of Som Hill) the covered relief tended to slope southward (probably due to the incision of Séd stream), so valleys could develop on the southern side. The valley between Felső-Hajag and Középső-Hajag descends from 610 m while at the north-western part of Felső-Hajag two valleys begin at 520. An inactive sink hole can be seen here. In the north-east the first valley starts at 545 m and further valleys range to the south with a sequence of valley heads of growing altitudes (550, 560, 570). Valley heads are found at 590 ad 580 m on the south-western and south-eastern sides of Rend-kő, respectively. The valley head between Rend-kő and Felső-Hajag lies at 570 m. Valleys did not develop on the western side of Középső-Hajag and Alsó-Hajag. Their heads are situated at 610 m and 590 m altitudes in the north. On the eastern side three inactive sink holes can be found towards the south. The valley in the eastern part of Középső-Hajag and its branch valleys joining in from Alsó-Hajag descend from 570 m.

Using the above described data the distribution of the burial of Hajag is shown in *Fig. 7*. The mass of Hajag has uplifted from the burial as several isolated elevations. These are the following: Felső-Hajag, the two elevations of Rend-kő and Középső and Alsó-Hajag.

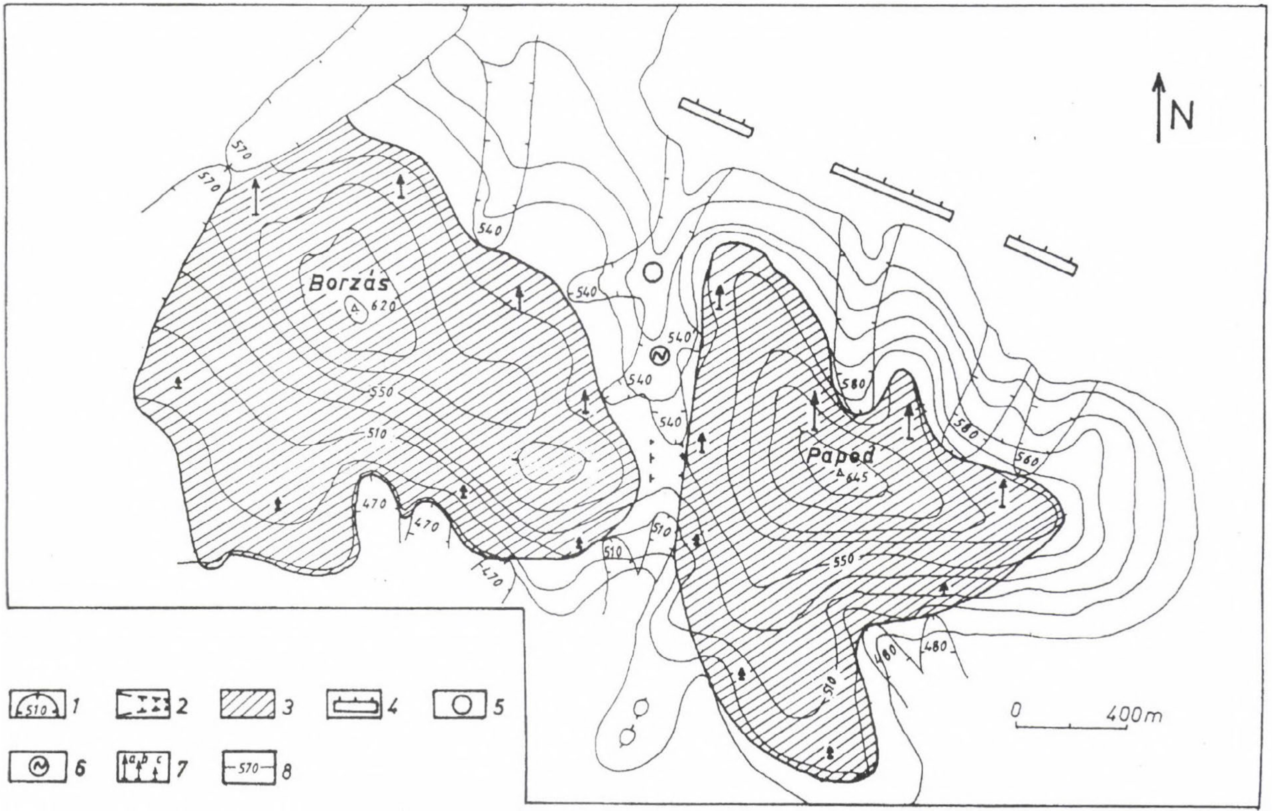

Fig. 6. Reconstructed map of the burial of Papod-Borzás. – 1 = valley head; 3 = valley with its former upper part; 3 = former uncovered area; 4 = fault; 5 = inactive sink hole; 6 = wallowing-place; 7 = degree of elevation: a) high; b) medium; c) low; 8 = contours

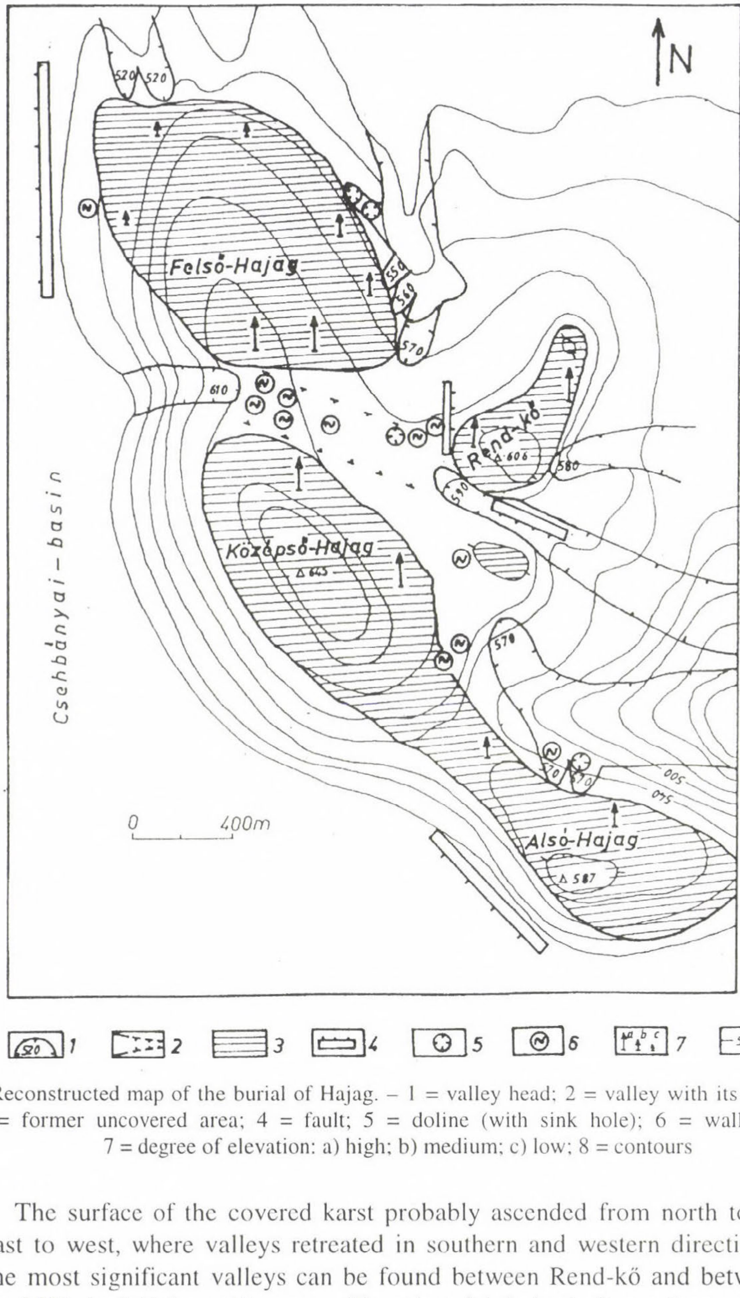

Fig. 7. Reconstructed map of the burial of Hajag. – 1 = valley head; 2 = valley with its former upper part; 3 = former uncovered area; 4 = fault; 5 = doline (with sink hole); 6 = wallowing place; 7 = degree of elevation: a) high; b) medium; c) low; 8 = contours

The surface of the covered karst probably ascended from north to south and from east to west, where valleys retreated in southern and western directions. Out of these the most significant valleys can be found between Rend-kő and between Felső-Hajag and Középső-Hajag with a row of inactive sink holes in its section over Rend-kő. The presence of inactive sink holes refers to the retreat of the rock boundary within the valley. On the other hand, inside the old valley west of the 610 m high saddle a younger valley sloping westward indicates its development after the denudation of the upper part of the older valley during an interval when the surface of the covered relief had already sloped westward. The western slopes of Hajag are poor in valleys. This refers to the fact that the relief in the Csehbánya Basin initially also declined from the north to the south. (Though the eastern slope of Hajag supposedly declined from the west to the east the valleys have gradually turned to the north indicating that in their wider surroundings the former covered relief probably declined to the north. Therefore the detection of the boundary of the former burial meets with certain difficulties.)

The rock boundary at the western margin of Hajag probably had had a north to south orientation. Here its location can be estimated using the analogy with elevation data of rock boundaries obtained from the other slopes of Hajag.

The case of probable uplift during and after exhumation

The difference of the altitudes of the valley heads makes it likely that after the development of the valleys the shift of the northern side of Som Hill and Kis-Som Hill from south-west to north-east and its southern side from Kis-Som Hill to north-east and south-west was decreasing.

The relative altitudes of Papod and Borzás –judging from the altitudes of valley heads – could be reversed compared to the present-day situation. That is the valley heads of Papod are located 30–40 meters higher than those of Borzás.

Therefore the rise of Papod (particularly its northern parts) exceeded that of Borzás by approximately 10–20 meters. (Currently Papod peaks at 645 m while Borzás is 620 m high.)

The valley heads on the northern side of the horsts are located 70–100 meters higher than those on the southern side. A conclusion can be drawn that the northern side of the horst group experienced a much more intense uplift than the southern one (transversal tilt during elevation).

Neither the mass of Hajag has had a uniform uplift. On the eastern side the degree of elevation increased from north to south toward Rend-kő and decreased from here towards Alsó-Hajag.

Considering the distribution of altitudes of the valley heads faults are presumed to exist within the horsts between Som Hill and Kis-Som Hill, between Borzás and Papod, Felső-Hajag and Középső-Hajag and between Felső-Hajag and Rend-kő. Geological maps also show faults at these locations (CSÁSZÁR, G. *et al.* 1981) which indirectly proves the correctness of the above described method.

Considering the highest valley heads of the horsts their relative uplift since the evolution of valleys can be estimated. Thus Som Hill elevated 40 meters higher compared with Papod, and 10 meters higher than Hajag and Hajag elevated 30 meters higher than Papod during the period in concern.

Summary

A method has been demonstrated which can be suitable for the determination of the borders of the former sediment cover. This requires a survey of the former rock boundaries. A map can be constructed based on the detection of the formations developed along rock boundaries. It can be carried out on karsts which are missing formations on their formerly uncovered parts (they were not developed at all or later became denuded) whereas they developed along the rock boundaries of another older sediment cover.

Such conditions could be present on a carbonate reliefs having seen peneplanation for a long period of time and subsequently elevated over the neighbouring terrain also built of carbonates. (This way these blocks neither were affected by the burial nor the current karstic water table lies near their surface. (Such topography is to be found in the Bakony Mountains.) A map of the former burial of the three studied horsts has been drawn. Analysing this map and the altitudes of formations developed along former rock boundaries the followings can be disclosed:

– The horst area has not uniformly uplifted from the burial but in several isolated spots. (In this way Som Hill and Papod-Borzás uplifted in two parts each while Hajag in four smaller parts). This means that the eminences of the former peneplain remained uncovered in the area of the examined horsts, or they did not exist in their present form during the burial period. In the latter case the horst groups obtained their present mountainous character during their young (Quaternary) uplift.

– The sediment cover could be the Csatka Gravel Formation. Since its surface sloped northward, in the surrounding of the horsts, probably it also had similar direction in the area bordered by the mountains (Hárskút Plateau, Pénzesgyőr area). This theory is in accordance with the opinion of experts in geology (KORPÁS, L. 1981).

– Directions of the sediment transport could partly change. Therefore, due to the denudation in the surrounding of Som Hill and Papod-Borzás the direction of sloping of the covered relief changed in the area south of the horsts. This could result in the development of a younger generation of epigenetic valleys.

– Horsts could have had a different rate of uplift so transversal tilt could take place in the Papod-Borzás area and longitudinal tilt might occur on Som Hill. Moreover, uplift of different extent could take place within the horsts themselves (between Som Hill and Kis-Som Hill, Borzás and Papod, Felső-Hajag and Középső-Hajag, Felső-Hajag and Rend-kő) but among the horsts as well.

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Á. KERTÉSZ and F. SCHWEITZER (eds): Physico-geographical Research in Hungary
Studies in Geography in Hungary 32, Geographical Research Inst. HAS, Budapest 2000, pp. 135–142.

EVALUATION OF SPELEOTHEM AGE DATA FROM BARADLA CAVE (NE-HUNGARY) WITH RESPECT TO LATE QUATERNARY CLIMATIC OSCILLATIONS

LÁSZLÓ ZÁMBÓ[1]–FORD, DEREK[2]–TAMÁS TELBISZ[1]

Abstract

Palaeoclimatic researches are of high importance, today. Since the karst processes are influenced by many climatic factors (mainly temperature and wetness), the intensity of karstification changes according to climatic oscillations. The speleothems enclose many information about palaeoenvironment. First of all their occurence suggests relatively humid and warm climate. In this paper, a speleothem growth intensity diagram is showed from U-series age data. The diagram is compared with the results of other investigations (malacological, palinological, loess-chronological, stable isotope and NW-European speleothem records) and a detailed analysis of Late Pleistocene climate is given.

Keywords: Speleothem, Late Pleistocene, Palaeoclimate

Introduction

Karst phenomena are highly impressed by changes in palaeoenvironment. Since climate is a very important and changing factor of karstification, the effect of climatic oscillations are well reflected in the intensity of karst processes. Thus, speleothems have special importance as palaeoclimatological records. Karstification is the most intensive in limestone areas covered by soil, because the soil has a crucial role in CO_2 production, which is the most important factor in limestone solution. Formation of soil cover was active in warmer interglacial and interstadial periods during Pleistocene. Lack of precipitation is also a limiting factor of karst corrosion, thus speleothem climate records have double meaning: intensive speleothem growth periods signify warm-humid conditions while limited growth intensity was dominant in cold or arid periods.

In order to reconstruct climatic changes of Central-Europe in Late Pleistocene, and for a better understanding of the development of Baradla Cave (NE-Hungary, *Fig. 1*), speleothem ages were examined by the U-series method. Baradla is the longest

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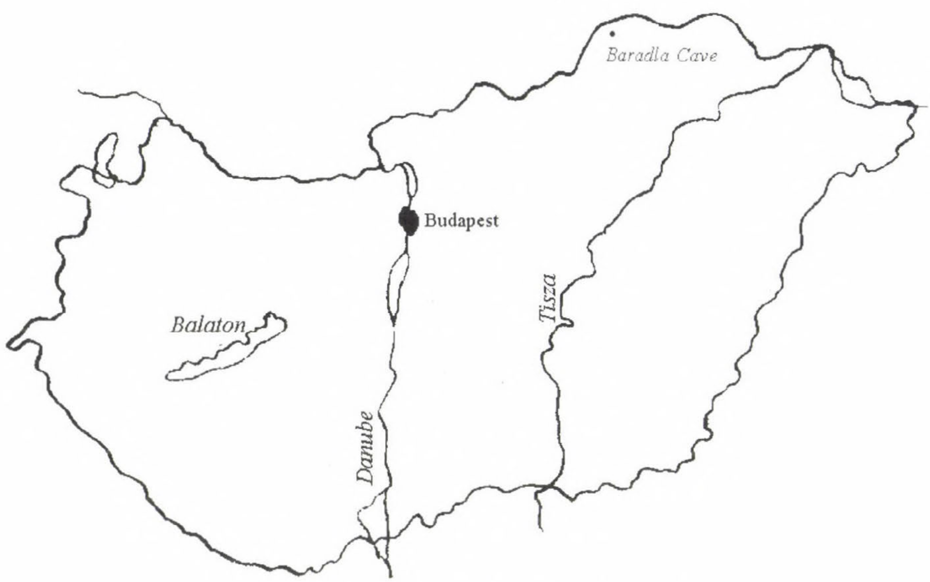

Fig. 1. Location of Baradla Cave in NE-Hungary

Cave in Hungary (with its total length of 24 km). This cave is an excellent example of the "Ideal Watertable Cave" (FORD, D.C.–WILLIAMS, P.W. 1989). This is an allogenic river cave created by streams flowing off of non-karst lowlands into limestone hill inliers. The main river passage now takes water only during floods, otherwise – in "normal" hidrological conditions – the speleogenesis is the dominant process.

Earlier researches have proved several erosional and sedimentational periods (PIROS, O. and GYURICZA, GY. 1986; SZENTES, GY. 1965) in Quaternary. Since the formation of lower, active river passage, there has not been significant widening of the main passage. Instead, occasional large floods have destroyed the sedimentation forms in the main passage and Late Pleistocene environmental changes have influenced the intensity of sedimentation. There are only few chronological data about the development of the Baradla Cave: LAURITZEN, S.E.–LEÉL-ÖSSY, SZ. (1994) determined U-series ages of some speleothem-bands (*Fig. 3*), but these data miss the error bars, so we could not use them in our analysis.

Our investigations aimed at the determination of the ages of older (probably oldest) speleothems in the main passage. The 17 samples were taken mainly from the basal part of toppled stalagmites and from flowstones on cave terraces. The ages were determined by U-series dating by the alpha spectrometry method. The speleothem samples from Baradla Cave have a low U-content and the $^{230}Th/^{232}Th$ ratio is small (probably because of great detrital contamination of samples by floods). These conditions reduce the accuracy of the measurements; however, these data can be used in a broad comparative analysis with the results of other methods (malacological records (SÜMEGI, P.–KROLOPP, E. 1995) and loess chronological data (PÉCSI, M. 1975, PÉCSI et al., 1979) from Hungary, stable isotope (^{18}O) studies (MARTINSON, D.G. et al., 1987), NW-European speleothem age data (BAKER, A. et al., 1993) and palinological analysis (GUIOT, J. et al., 1989). The results from Baradla Cave show strikingly good correlation with other time series, which makes it possible to infer Late Pleistocene palaeoenvironmental changes based on speleothem record.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 022977.

Results

A. Development of Baradla Cave (some conclusions)

– None of the deposits (not even the oldest parts of the biggest toppled columns) is likely to be as old as 200,000 years BP. The formation of the cave probably began before the Late Pleistocene, but in the last large erosional period, the remains of older deposits had to be destroyed.

– The main river passage is likely to be at least 150,000 years old.

– The dimensions of the bedrock passage have not changed very much since the last interglacial.

– The ages of the terrace deposits suggest that the cave was more aggraded during the Würm than it is today. Floods that built gravel and clay deposits up as much as 4.5 m above the modern stream bed were probably also responsible for toppling the very large columns.

B. Palaeoenvironment changes during Late Pleistocene

Growth frequency graph can be drawn from speleothem age data. For this purpose, 2 steps were needed:

a, Instead of preparing a simple histogram from speleothem age data, the "point-like" age data were replaced by a simplified probability distribution function counted from age and σ errors. Finally, these functions were summed up for getting *Fig. 2*, which is more informative than a histogram.

b, Since the frequency values are changed in step a, all values have to be divided by the mean value of the period so a %-scale can be adapted for the Y-axis.

Results were compared with other time series displayed in *Fig. 3–6*. The age limits of our measurements are from 180 ka BP to 10 ka BP. The younger limit comes from the examination goal (i.e. the sampling principally aimed at the collection of older speleothems from the main passage), and does not mean the lack of Holocene speleothem growth in Baradla Cave. This period involves the end of Riss glacial, the Riss-Würm interglacial and Würm glacial. Some peaks and troughs on the diagram can be observed, these are marked by Roman numbers. The climate periods are examined according to this diagram from the beginning of the age data.

At the end of Riss glacial, limited speleothem growth and cold temperature (*Fig. 5*) were characteristic in NW-Europe and in Hungary, too (cc. to 125,000 BP).

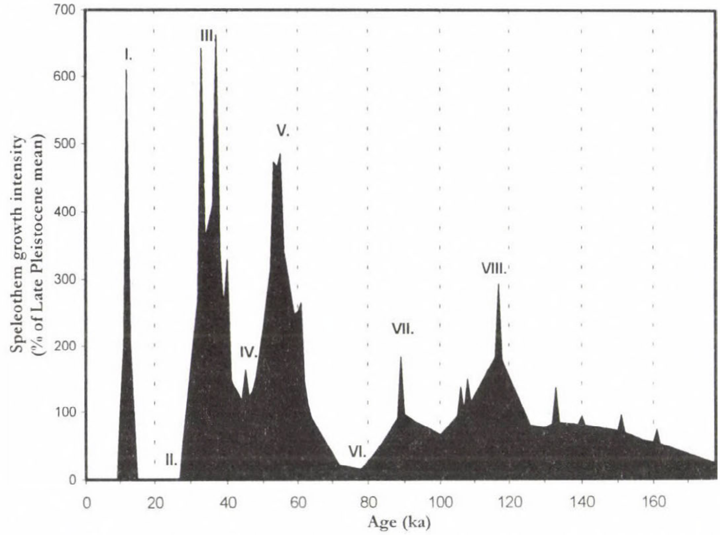

Fig. 2. Speleothem growth intensity based on speleothem age data from Baradla Cave

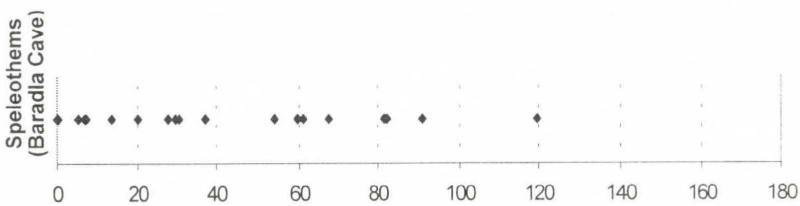

Fig. 3. Speleothem age data from Baradla Cave (edited from data in LAURITZEN, S.E.–LEÉLÖSSY, Sz. 1994)

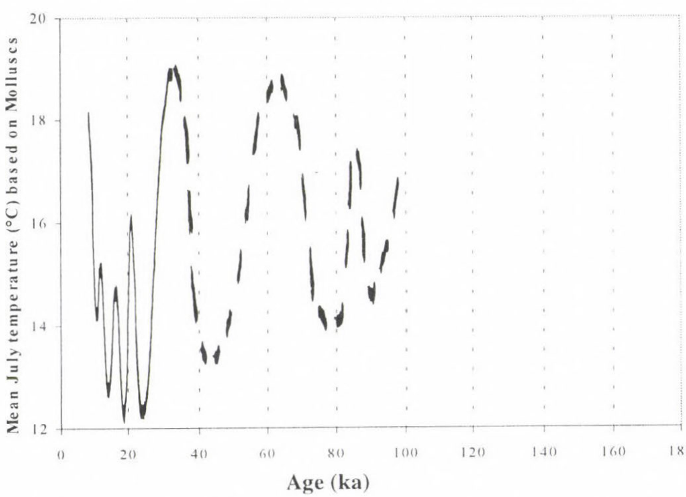

Fig. 4. Climate oscillations based on Molluscs (SÜMEGHY, P.–KROLOPP, E. 1995)

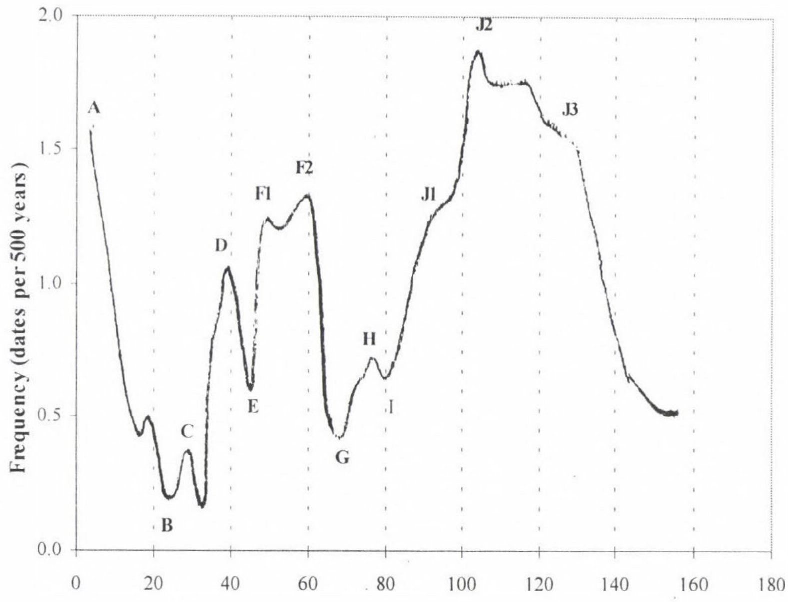

Fig. 5. Speleothem growth intensity based on speleothem age data from NW-Europe (BAKER, A. et al., 1993)

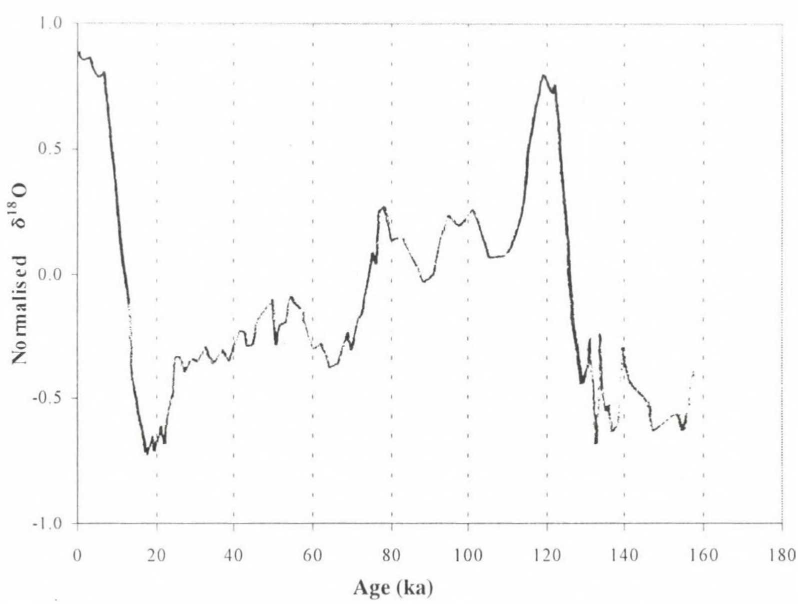

Fig. 6. Oxygen isotope record (MARTINSON, D.G. et al., 1987)

The Riss-Würm interglacial can be marked by the cyclicity of climate: 3 interstadials (Eemian: 125 ka – 115 ka; Brørup: 105 ka – 95 ka ; Odderade: 85 ka –75 ka) are distinguished based on ^{18}O-isotope data. The Eemian (the warmest?) is recognisable on Hungarian and NW-European speleothem growth diagrams (VIII., J3), but karst records of later interstadials seem to be shifted in time (VII., J1: 90 ka BP) and in Odderade interstadial the speleothem growth intensity is very low (VI.; I,H,G). Excluding chronological errors, BAKER A. et al. (1993) concluded that this situation was the consequence of aridity during this relatively warm interstadial. At the end of this interstadial, the Fennoscandian ice sheet started to grow causing a high pressure system and aridity. There are other proofs for arid conditions in that period such as wind-blown sands in British caves and loess deposits from England. Pollen records from France (GUIOT et al., 1989) suggest that the Odderade interstadial was only slightly drier, but was preceded and followed by very arid intervals.

There was a strong deterioration of climate in Early Würm (VI.; G), characterized by loess-formation in Hungary and glacial deposits in Denmark and Poland (TL ages: 60–80 ka; KRONBERG, C. and MEJDAHL, V. 1990). In the würm$_{1-2}$ interstadial (60–50 ka) the climate was warmer, and relatively dry (according to SÜMEGI, P. –KROLOPP, E. 1995) based on Molluscs (with steppe vegetation). The enhanced speleothem growth (V.; F2–F1) contradict this opinion, because it suggest more humid climate. (The palinological data show arid climate at the beginning and humid at the end of this interstadial.) A possible explanation is that a greater humidity gradient existed between the Carpathians and the enclosed basin climate than generally; however this problem requires further works.

Except for the O-isotope record, all diagrams show the cold climate of Midlle Würm (IV.; E; pollen and molluscs; 45 ka BP) and the warm Würm$_{2-3}$ interstadial (III.; 40 ka), though in the duration of this period there are some differences. In NW-Europe the cooling down began 35 ka BP (C), marked by discontinous permafrost and tundra vegetation at N from Netherlands. Meanwhile in the Carpathian basin the soil formation was working on. (Mende Upper Soil, 32–27 ka BP, PÉCSI et al., 1979) and enhanced speleothem growth was dominant. In this period, a strong N–S climatic gradient was likely to determine the European climate, because there are other proofs for relatively warm climate in southerly Europe (GUIOT et al., 1989; ROUSSEAU, D.D–PUISSÉGUR, J.J. 1990).

Late Würm brought extremely dry (SÜMEGI, P.–KROLOPP, E. 1995) and cold conditions in the Carpathian basin from 27 ka BP and this caused the halting of speleothem growth (II.). The loess sedimentation was the dominant process, though interrupted by a short warmer event (O-isotope data; Soil formation: 22–20 ka, PÉCSI, M. 1975), but there is no evidence for intense karstification from this time (Probably, because of the short duration.).

There is a relatively rapid warming period from the end of Würm to now. This led to the disapperance of cold tolerant mollusc species, the recommencement of soil formation and forestation, thus, the intensification of karst processes (I.).

Conclusions

It is concluded that speleothem growth is a good indicator of climate changes, and sensitive both for temperature and humidity. This makes it necessary to use this speleothem records completed by other methods in the reconstrucion of palaeoenvironment. There are several possibilities to improve the efficiency of speleothem examinations, e.g. detailed study of speleothem rings (stable isotopes: ^{18}O, ^{13}C (HENDY, C. 1971; BROOK, G.A. 1990; VESICA, P.L. et al., 2000); organic acid content (MCGARRY, S.F.–BAKER, A. 2000), which is the objecitve of further invetigations.

The comparison of data from NW-Europe and the Carpathian basin has showed that climatic oscillations are more or less parallel but there are some differences which can lead to interesting conclusions.

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LANDSLIDE ACTIVITY AND LAND UTILISATION AT THE HIGH RIVER BANK ZONES*

JÓZSEF SZABÓ[1]

Abstract

The paper focuses on the currently most active landslide area in Hungary, the valley of river Hernád. It reviews the level of the landslide activity and its fluctuation over time based on detailed landslide and land utilisation mappings as well as their relation to land utilisation in the high bank riverside zones. The cultivation is dependent on several social conditions, that have a uniform impact on this relatively small area. However, the territorial changes of the cultivation and the decline of the quality of gardens is not accidental, but concentrated on certain areas within this high bank zone. The conclusion of the research is that in this section of the Hernád valley the main aim should not be to prevent landslides on the most endangered bluffs. Instead, these areas should be reserved in its natural status with unique geomorphologic processes and forms by granting protection and presented to the public.

Introduction

In Hungary, the majority of landslides and landforms of sliding origin occur in three different types of landscape:

1) The stable forms of the previous massive but usually inactive movements occur at the marginal zones of the Neogene (strato)volcanic mountains as significant features of the landforms.

2) In the slope evolution of the hilly regions built up of unconsolidated, for the most part Neogene sediment, landslides had a special importance in several phases of the Quaternary, and these processes are still in effect in many areas.

3) Mass movements, and especially landslides are formative processes in the geomorphologic development of the bluffs along several rivers in Hungary (Danube, Rába, *Hernád*).

In many of his previous studies focusing on different landscapes, the author gave a survey about the sliding processes in the different types of landscapes playing a decisive role in geomorphic development (e.g. SZABÓ, J. 1985, 1995, SZABÓ, J.–FÉLEGYHÁZI, E. 1997). A general summary of the topic was published in an independent volume (SZABÓ, J. 1996). The present paper focuses on the relationship between the movements and economic utilisation in the currently most active landslide

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movements and economic utilisation in the currently most active landslide area in Hungary, the high bank area of the Hernád river. Similar problems were surveyed earlier in different sections along the Danube (lately among others: PÉCSI, M., SCHWEITZER, F. and SCHEUER, Gy. 1987, PÉCSI, M. 1994, LÓCZY, D., BALOGH, J. and RINGER, Á. 1989).

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 014948.

Discussion

General features of the Hernád valley and of the high bank development

The Hungarian section of the Hernád valley is situated along one of the most important tectonic line of the Carpathian Basin (Hernád line). However, its morphological features are basically the result of the processes induced by erosional slope evolution. As a consequence, half a century ago several terrace levels have been mapped above the present valley (LÁNG, S. 1947). Nevertheless, the occurrence of terraces is rather haphazard. They are usually missing on the sliding sections of bluffs.

The valley has incised into the sediments of the Pannonian Sea (Upper Miocene), its high banks expose Pannonian sediments of very poor stability covered by a loess layer of maximum 5 meters thickness for a longer stretch. Pannonian sediments are represented by silty-clayey deposits and within them horizons affected by oxidation and reduction (of yellowish-brown and grey colours, respectively), and alternating with them sandy layers. Accordingly, the degree of impermeability by water also frequently fluctuates within a given vertical profile which is a decisive factor with respect to sliding.

High banks occur at three sections of the Hungarian part of the Hernád valley *(Fig. 1)*. The *southern high river bank* is about 30 km long, the steep slope zone is 500–700 m (maximum 1250 m) wide on the average, and its elevation above the valley floor is generally 70–80 m (with 20 m minimum and 145 m maximum). The freely meandering river is currently undercutting the bank at 18 places in the area of the slope stretch (18 km^2). Undercutting increases the steepness of the bank, and due to the lateral erosion its height also grows. Therefore, it also controls the mass movement processes. The model created on the basis of the local surveys (SZABÓ, J. 1995). As a rule initially there are falls on the bluffs which become higher due to undercutting, then these (usually in the upper zone) are followed by *landslides*. Since the bends are gradually sliding downward in the foreground of the bluff, the places being undercut are also constantly changing. Thus, the parts of the slope in active sliding are also changing gradually. The most active slopes occur at places abandoned by the meander after a longer period of undercutting. Following the cessation of undercutting, the inner "reserve energy" of the slopes sustains the movements for a longer time, although their intensity decreases depending on the climatic conditions and due to the continuous decrease of the slope angles. As a result of the depletion of the inner reserves, undercutting by a new meander may start, and the mass movements may be reactivised.

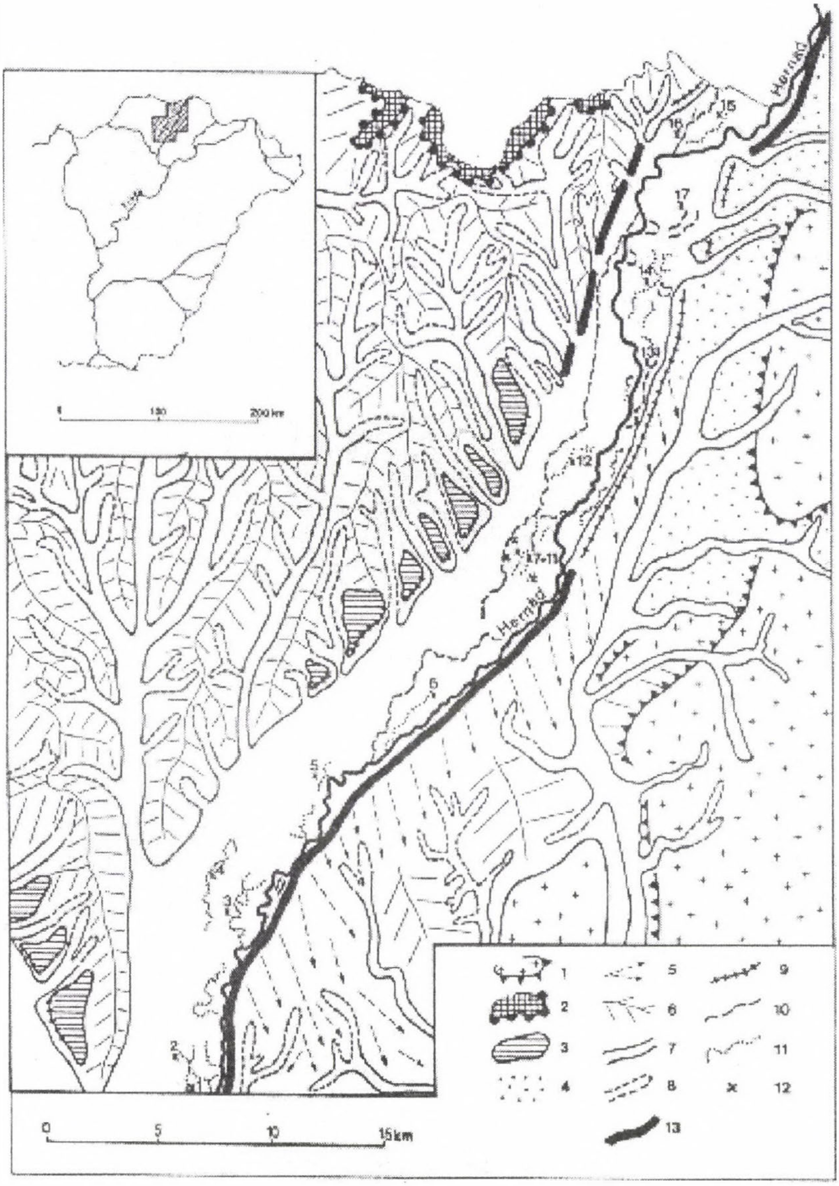

Fig. 1. Geomorphological position of the Hernád valley. – 1 = area of volcanic mountains with marginal terrace of structural origin; 2 = remains of the Cserehát glacis surface, covered by Early Pleistocene gravel; 3 = glacis surfaces in Pleistocene valleys; 4 = pediment surfaces in the foreland of the Zemplén Mountains; 5 = general slope direction on the left bank of the Hernád; 6 = slopes undistinguished; 7 = erosional valleys; 8 = derasional valleys; 9 = terraces along the Hernád; 10 = remains of abandoned riverbeds in good state in the Hernád valley; 11 = strongly accumulated destroyed remains of riverbeds; 12 = site of sampling for pollen analysis; 13 = high banks

The surveys show that this process may have been repeated several times during the Holocene, thus the recession of the front of the high bank was significant. The rapid retreat is primarily responsible for the fact that practically no subsequent valleys could develop on this high bank reach, and the terraces are also missing.

Landslide activity and hazard

In the course of the mapping of the landslide forms of the high banks it became clear, that smaller sectors on the demonstrated slope stretch where the unambiguous identification of the landslide forms is possible, occur only exceptionally. The slope is generally characterised by a contiguous chain of slide formation groups of different age. Therefore, the unanimous separation of the different slides is not only problematic but in most casesit is not necessary as well. It would be much more reasonable to determine what the general state of the landslide forms is like along the part of the slope in concern, and in this way the activity of the slide could be determined. Aiming at this, the individual sectors of the high bank slopes were grouped into five grades of landslide activity.

grade 0: no traces of slides can be identified unambiguously;

grade 1: the slope has been stable for a longer period (hundreds of years), the remnants of the slides can only be seen in the forms of slight undulations;

grade 2: temporarily stabilised slopes with easily identifiable landslide forms. There are closed depressions between the slipped masses;

grade 3: open (usually younger than 10 years) rifts show that the movements of the slope renew from time to time;

grade 4: currently sliding slope stretches, or practically annually reoccurring slumps.

The comprehensive band chart of the results received on the 20 km long stretch along the southern bluff can be examined on *Fig. 2*. It is apparent that in the revival periods of the sliding (usually during the months following wet winters) the area of the active parts increases with the regression of the scarps. This process can be detected not only in the lower part of the slope but also in the middle zone with previous slides. During these periods, the uppermost scarps of the landslides usually are also retreating, thus slides occur on new areas as well.

The variety of the landslide forms on a shorter stretch of the surveyed area and the spatial distribution of the activity grades are illustrated in *Fig. 3*.

Description of the land utilisation in the high bank zones

The Hernád valley used to be a traditionally important trade route between the inner areas of the Carpathian basin and the surroundings of the Carpathian Mountains and the land located north of it as well. However, the settlements along the valley are basically agricultural. This is particularly valid for the studied high bank zone. There are 10, so called tiny villages with a few hundreds of inhabitants each on the 30 km long

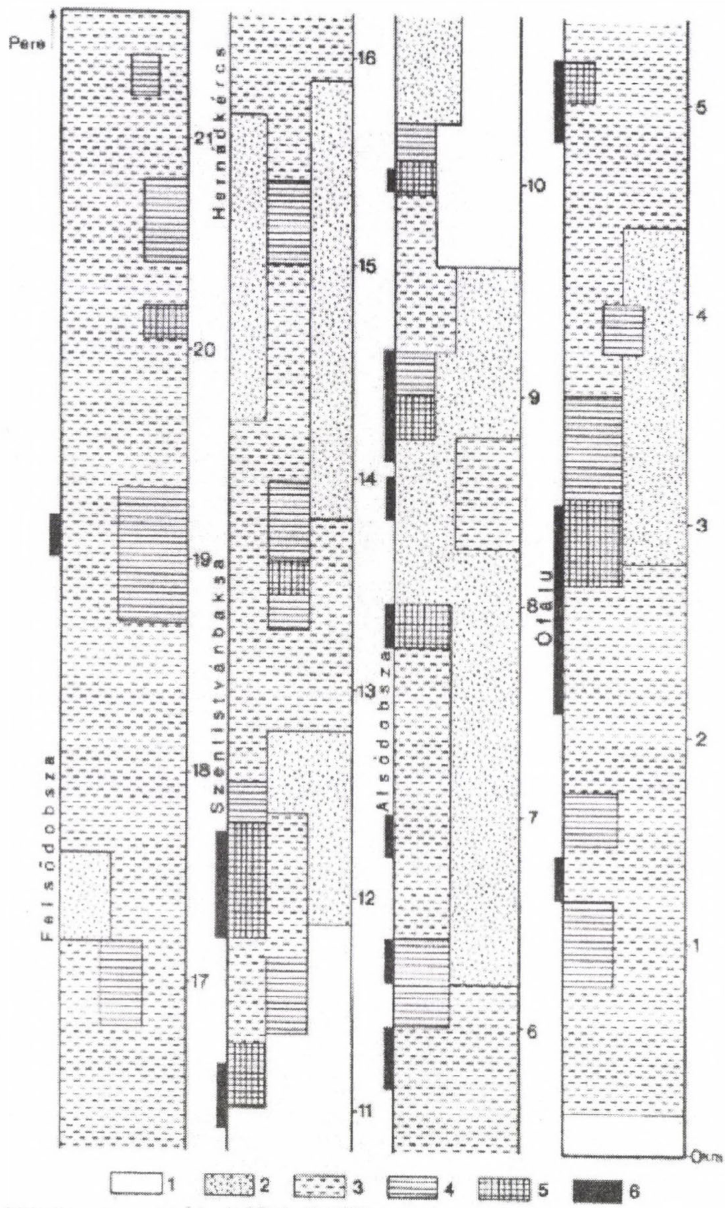

Fig. 2. Activity grades of slide slope evolution on the left bank of the Hernád between Pere and Gesztely (0–22 km). – 1 = slopes without any forms of landslide (activity grade 0); 1 = stretch of bank characterised by strongly degraded forms (activity grade 1); 3 = parts of a slope with recent slide forms, now inactive (activity grade 2); 4 = recent slide slope with active parts (activity grade 3); 5 = part of slope characterised overwhelmingly by active slides (activity grade 4); 6 = the river is currently undercutting the lower part of the high bank

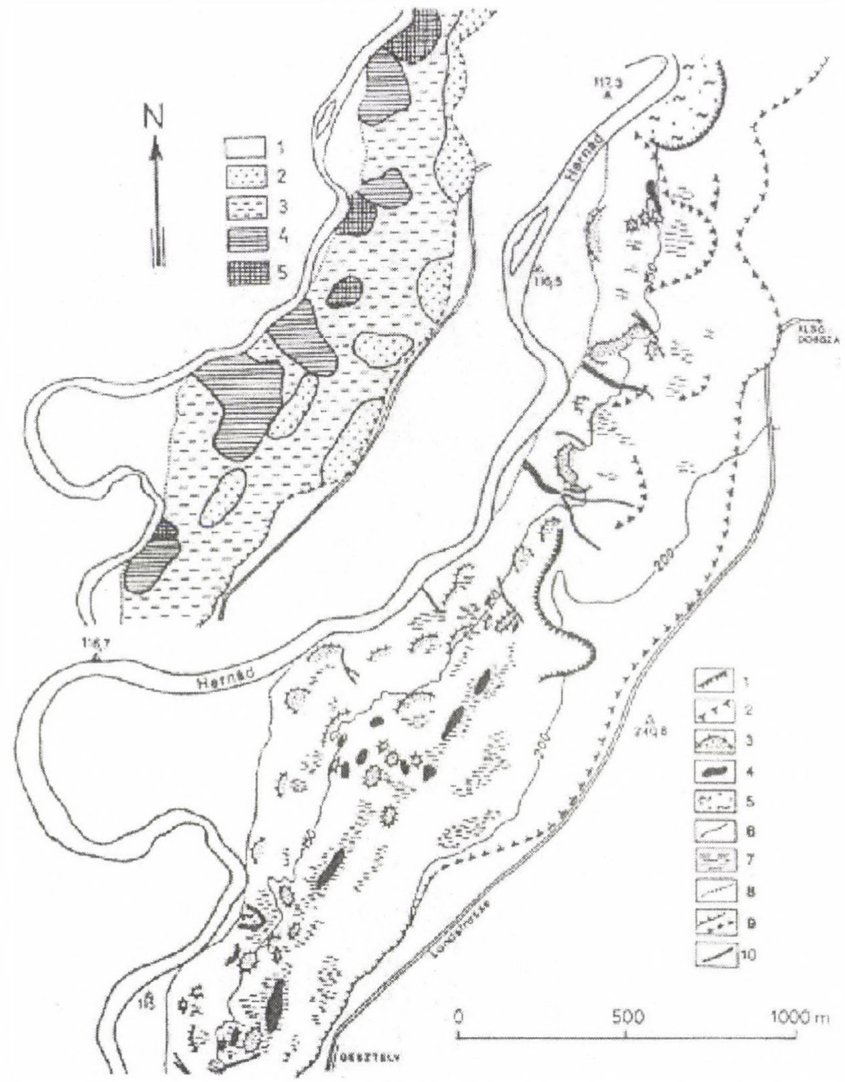

Fig. 3. Slide forms and slide activity on the Hernád high bank between Alsódobsza and Sóstófalva. Main map: – 1 = fresh scarp without plant cover; 2 = degrading scarp; 3 = larger slid masses with marked edges; 4 = closed, marshy-boggy depressions; 5 = active slide surfaces without plant cover; 6 = contour lines (in m a.s.l.); 7 = mostly closed depressions; 8 = lower edge of the high bank; 9 = upper (in some places, wall-like) edge of the high-bank zone; 10 = young erosional furrows. Insert map: – 1 = activity grade 0, 2 = activity grade 1, 3 = activity grade 2, 4 = activity grade 3, 5 = activity grade 4 (situation at the end of the 1990's)

stretch. No industrial factories operate in these settlements, and the inhabitants either commute to work to the neighbouring industrial area – which is currently in crisis –, or mainly deal with farming. The overwhelming part of the cultivated land can be found on the gently sloping hilly part in the eastern background of the high river bank. Arable land is the typical land utilisation form. The proportion of arable land is inferior on the high bank slopes *(Fig. 4)*. Here, pastures are in prevalence, and the share of woodland is relatively high compared to the surrounding area. Unfortunately, a significant part of them is planted forest of very poor quality which can be hardly utilised today. This is equally the consequence of the landslides and nowadays of mismanagement. As a result of the *ad hoc* forest clearings and the recently the decline of animal grazing, a significant part of the forests and pastures indicated on the map are today bushy-scrubby areas. That is, they are hardly utilised.

It is remarkable that gardens (orchards, vineyards) are relatively extensive on the slopes of the bluffs. In the zone shown on *Fig. 3,* 19.4% of the total area belongs to this category. Field mapping of a scale of 1:5000 relying on the topographic maps made it obvious that currently even the intensive horticulture experiences decline over the area. The reasons are partly social. On the one hand, there is a serious loss of population in the neighbouring villages; on the other hand the trend characteristic during the seventies and eighties (that many people living in the neighbouring Borsod industrial area bought gardens and cultivated land in this area) has practically stopped.

Table 1. State of the gardens on the bluff of the Hernád between Sóstófalva and Alsódobsza (the section of Fig. 2. between 4.5–8 km) according to the location on the slope

Degree of cultivation		Upper part		Lower part	
		area (ha)	%	area (ha)	%
I	very well kept	3,7	10,8	0,7	7,0
II	well kept	1,6	4,7	1,0	10,0
III	cultivated	2,1	6,1	0,1	0,1
IV	unkept	3,5	10,2	1,0	10,0
V	abandoned	5,8	16,2	1,8	18,0
VI	traces of horticulture	17,6	51,3	5,4	54,0
Total:		*34,3*	*100,0*	*10,0*	*100,0*

When mapping their spatial pattern, the gardens were classified into six groups (very well kept, well kept, cultivated, unkept, abandoned, with traces of horticulture) with regard to their general condition. As it is indicated by *Table 1* the distribution of gardens on the slope is not uniform. Only one-fifth of the ca 45 hectares of gardens are situated in the lower part of the high bank zone. The gardens are generally hardly looked after and the proportion of abandoned, and almost unidentifiable vineyards and orchards is especially high in the lower part of the river bank. These categories constitute almost three-quarters of the gardens in that section.

Comparing the above mentioned proportions with the territorial extension of each landslide activity grade and with its situation on the slope, a clear-cut relationship can be seen between the intensity of horticulture and the landslide hazard (or activity).

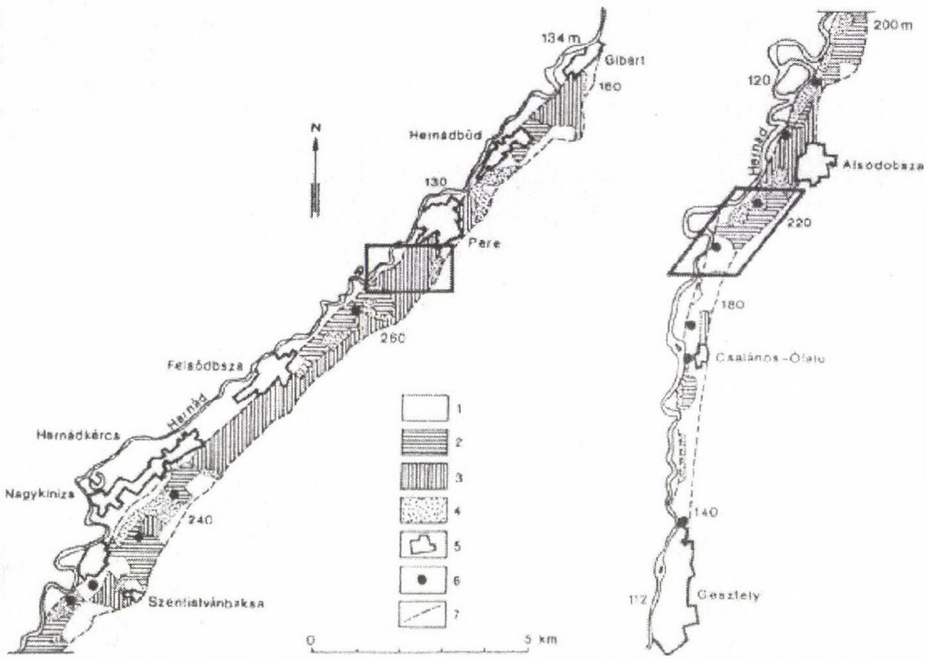

Fig. 4. Land use and occurrence of slides along the southern section of the high bank zone of the Hernád. – 1 = arable land; 2 = garden, orchard, vineyard; 3 = pasture; 4 = forest; 5 = settlement; 6 = more important slides; 7 = boundary of the high bank zone. The areas proposed for conservation are contoured with thick lines

In the area shown on *Fig. 3* (Sóstófalva), the areal extension of the individual activity grades can be characterised approximately by the following proportions: grade 0: about 5–10%, grade 1: 15–20%, grade 2: 45–50%, grade 3: 15–20%, grade 4: 4–5%. The slope parts belonging to grade 4, and 80% of those belonging to grade 3 are located in the lower part of the slope. The small number of gardens which are currently abandoned indicate that the proportion of areas under horticulture has always been smaller in these areas because of the frequency of sliding. In addition to this, nowadays – when the abandonment of gardens became a characteristic trend –, the owners have cancelled the cultivation of the lower gardens seriously affected by landslide hazard.

Conclusion

There is not a real chance for preventing the high bluff landslides caused by the lateral erosion of the river in the Hernád valley. A complete regulation of the river – through the elimination of the meander development – would only ensure the fixing of slides only in a long run, after the exhaustion of the stress within the slopes. In the near future, the slides will renew from time to time depending on the climatic conditions, and their stabilisation cannot be economically efficient even with high investments because of the slip planes lying chiefly at a 5–10 meters depth. At the same time, a significant proportion of the high river bank is poorly utilised and it shows a definite decreasing trend.

In compliance with the above circumstances, the author claims that on an extensive stretch along the high banks of the Hernád river there is no sense to make an attempt to prevent mass movements. Even so, the main objective should be to conserve the slopes which anyway mean only a low and uncertain income for its owners. This open-air geomorphologic laboratory (the active slide evolution triggered by undercutting of the bluff by the river) which can even be treated as a natural curiosity. It might sound an unusual objective to ensure the free operation of the landslide movements but the processes and forms which can be seen here are instructive and picturesque. They could be presented for the public if it protection was granted. It is especially valid for the stretch (described in the present paper) with an otherwise minimal agricultural utilisation which would not be disturbed by the renewing landslides, and where the creation of a study trail could offer a source of revenue through the visitors.

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Dr. János Kubassek

Hungarian Hermit of the Himalayas

The life of Sándor Kőrösi Csoma

Dr. János Kubassek's travel books and biographies are always a success, with a very simple reason: sound learning, a wealth of personal experience, worthwhile subjects, a style which manages to be both passionate and natural, and a clear understanding of geography, history, culture and human life itself. Kubassek, de director of the Hungarian Geographical Museum in Érd, near Budapest, has produced this new work, a biography of Sándor Kőrösi Csoma, not simply a monograph of a fascinating character, but also an act of faith. Since his student days Kubassek (born in 1957) has been following Csoma's trail, first to Transylvania, where he was born, and later inspired both by his university studies and the work of other men who has fallen under Csoma's spell (Ervin Baktay, Dénes Balázs, Ödön Jakabos and others), he set out in 1980-81, founded by his own modest resources, to trace a path of this "patron saint on Hungarian orientalists" all the way to Darjeeling. Kubassek's journey took him to some of the remotest and least-accessible parts of the world, to the mountain fastness of the western Himalaya, Zanskar and Ladakh. In 1998 Kubassek returned to Western Tibet, and accompanied by a single local guide and three horses, once again sought out monasteries of Zangla and Phuktal, where Csoma had once lived and studied. Over the past two decades, Dr. Kubassek has given many hundreds of lectures on Kőrösi Csoma, the most famous Hungarian traveller of all time.

Hungarian Ancient History Research and Publishing Company
Budapest, 1999

Á. KERTÉSZ and F. SCHWEITZER (eds): Physico-geographical Research in Hungary
Studies in Geography in Hungary 32, Geographical Research Inst. HAS, Budapest 2000, pp. 153–159.

MAN-MADE IMPACT ASSESSMENT AT GEOTOPOLOGICAL LEVEL

SÁNDOR MAROSI[1]

Introduction

The effects of the considerably expanding human activities manifest in the geographical sphere could be examined and assessed by several methods and by investigations carried out in different scales (varying in details). In this respect a great number of examples present themselves in Hungary, including projects run in the Geographical Research Institute of the Hungarian Academy of Sciences, and the author's own research, too. Of them one of the more than 20 so-called "representative investigations of type regions" is presented below. It was carried out within the frame of the landscape research in the microregions in topological dimensions. This region is located in the flood-plain of the Danube, south of Budapest, in the southern part of the Csepel Island, in the vicinity of the settlements Lórév, Makád and Szigetbecse (MAROSI, S. 1953; GÓCZÁN, L. et al. 1973; MAROSI et al. 1973; MAROSI, S. and PAPP, S. 1978).

Description of the studied area

The test area is part of the relatively uniform alluvial plain of the Danube (the relative relief does not exceed 4–5 m). Due to its micromosaic properties it offers an excellent opportunity for the identification and classification of flood-plain facies and for the assessment of impacts exerted on the ecological factors by anthropogenic activities, on the basis of complex agroecological and bioecological surveys and measurements.

West of the built-up area of Makád, starting from the almost circular cut-off ox-bow lake complex lithological, morphometrical, pedological, phytogeographical, hydrological and microclimatological surveys were performed and measurements carried out in a 1 km long north to south section including the following types: 1. reed-beds (94.4 m a.s.l.), 2. fringe of reed-bed with high sedge (94.5 m), 3. hayfield (94.7 m), 4.

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willow-poplar shrub (95.0 m), 5. arable land on the ox-bow spur without vegetation (96.2 m), 6. arable land on the floor of filled ox-bow lake without vegetation (95.8 m), 7. arable land on the south slope of the elevations between channels without.vegetation (96.5 m) and 8. the maize field of the top surface between filled channels (98.4 m). These allowed the comparison of biotopes described by riverside vegetation still conforming to the natural succession with the cultivated agroecotopes of slightly different exposure and altitude.

Limited relief is not only characteristic of the microclimatic profiles, but of the whole test area surveyed and mapped in detail agroecologically (*Fig. 1*). However, on the surface minutely dissected by channel remnants a multitude of short but steep slopes is also typical. As a consequence, cultivation produced relatively extensive zones of "earthy barrens", indication of the erosion of the initial soils. Apart from them, erosion has affected other soils to different extent. On the former channel floors fresh accumulation of semipedolites is typical. Along the point-bars accompanying the riverbanks built of loose sand not covered by loessy silt (a common deposit in other places) and under cultivation sheet-wash and partly deflation also induced considerable soil erosion. The same applies for closed elevations.

Over the period since flood control measures the impact of the Danube on the ecological conditions and land cultivation has been mostly indirect and manifest through the control of groundwater conditions. Groundwater level is relatively high and has a cosiderably wide range of fluctuation. The series of soil types from (the driest) "pseudomicelian" chernozem to (the wettest) boggy meadow soils reflects subsurface moisture precisely. Minimal differences in relief (some tens of centimeters) and in the depth of groundwater table below surface allow the development of a distinct soil variety reflecting different ecological conditions.

The terrain at 97–98 m above sea level, classed as the higher flood-plain, has long been providing suitable conditions for the formation of "pseudomicelian" chernozems. (Before the flood control and water regulation of the Danube, the river extended over larger areas and it had lower water levels during inundation.) On the lower flood-plain levels at 94–96 m (particularly on the lower-lying sections) semihydromorphous or even hydromorphous soil formation was typical. An opposite process of soil desiccation is produced by cultivation with agrotechnical intervention and aeration.

Actual and proposed land uses

Two kinds of human intervention contributed considerably to the formation and transformation of the present landscape in general, and of the topological units in particular: 1) flood control measures and 2) the advent of farming. Both interventions suppressed the processes of hydromorphous and semihydromorphous soil formation considerably.

After the detailed evaluation of data, the generalisation of results suggest the following conclusions and proposals concerning land utilisation.

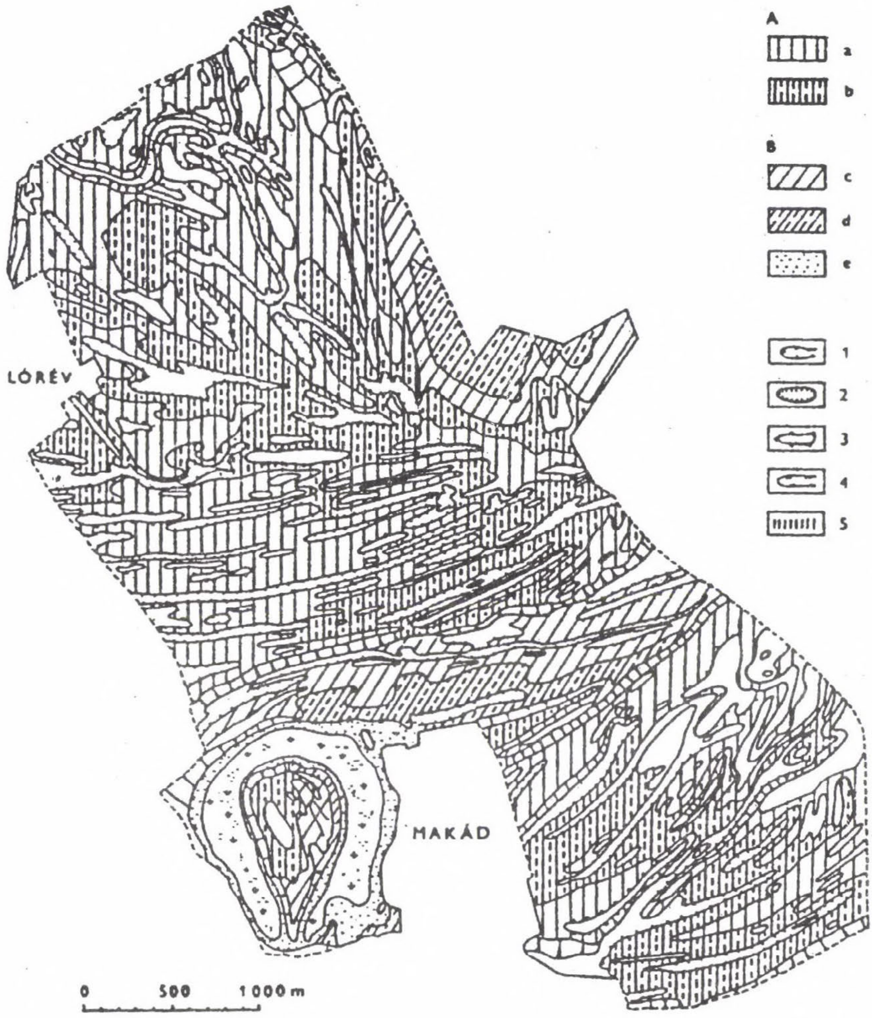

Fig. 1. Geomorphological map of the Danube valley floodplain (section between Lórév and Makád) (after GÓCZÁN, L., MAROSI, S., PAPP, S. and SZILÁRD, J.) – A = High flood plain: a = first level (ca 98 m a.s.l.); b = second level (ca 97 m a.s.l.). B = Low flood plain: a = first level (ca 96 m a.s.l.); b = second level (ca 95 m a.s.l.); c = third level (less than 95 m a.s.l.). 1 = cut-off meander filled with sediments; 2 = cut-off meander identifiable in the terrain; 3 = minor elevation; 4 = ox-bow lake; 5 = slopes

Among the identified 8 topological units investigated in details (*Figs 2* and *3*), those described by closed and high-grown plant stands were distinct primarily from those of open and low-grown vegetation. Further opportunities for the differentiation were provided by measurements at locations covered by natural and cultivated plants within the closed stands. A further refinement of ecological description might be achieved on the basis of soil and air humidity deriving from the higher or lower position. The basis of the differentiation between the open ecotopes is primarily of geomorphological character (lower or higher situation, exposure) and, in connection with it, of the hydrogeographical type (groundwater level) difference.

As ecological investigations confirmed by microclimatic measurements indicate, beside the different physical factors and endowments, and even counteracting them, the different functions of physical factors are complemented by human interventions, which create entirely new environmental conditions, agrogenic units, agroecotopes replacing the ecologically natural ecotopes. In our case among the cultivated surfaces the hayfield stock partly reflects this. The most typical agroecotopes are the units (limited to 3 microclimatic stations) observed on ploughland where among the physical factors the effects of the topography, lithology, groundwater table and (as a result of their interaction), the soil only observable against the background of the agrogenic factor prominent at present exerting mainly homogenising impact during the time of the measurements (ploughing, fertilisation, chemization etc.).

A contrast to the above three stations is the maize field with a characteristic stand climate representing a different agroecological unit.

The results of complex investigations combined with microclimatic observations attest to the adjustment of land use to ecological endowments. The ox-bow lake with open water surface (no measurements were carried out here) is suitable for fish or wildfowl breeding. The preservation of the reed-beds is motivated, in addition to economic considerations, by the excellent habitat it provides (high stand). The next in succession are the reed margin, hay meadow, willow-poplar bush which are not appropriate for cultivation since they are partially and seasonally waterlogged and have high groundwater tables. At the individual sites of measurements the limited extension of biotopes do not call for expansive amelioration measures, along other sections of the now flood-free alluvial plains of the Danube (which in total constitute a considerable area), other forms of land utilisation can also be envisaged, depending on the depth of groundwater table. For longer periods the actual level of the Danube controls groundwater and excess water conditions, which has to be taken into account in the case of cultivation. It is primarily the soils that can be decisive for the perspectives of utilisation of agroecotopes mainly through hydrological factors. In this respect the profiles of stations 5 through 8 indicate that the ecological properties of the mentioned ecotopes allow rather diverse agricultural cultivation. Among them the lowest situated station, No 6, with the highest ground water table has seasonal excess water hazard.

Locally the depth of groundwater creates suitable conditions for vegetable gardening, particularly in the drier biotopes and those in higher position, with willow-poplar bush in the present succession. No doubt, however, that particularly in the case

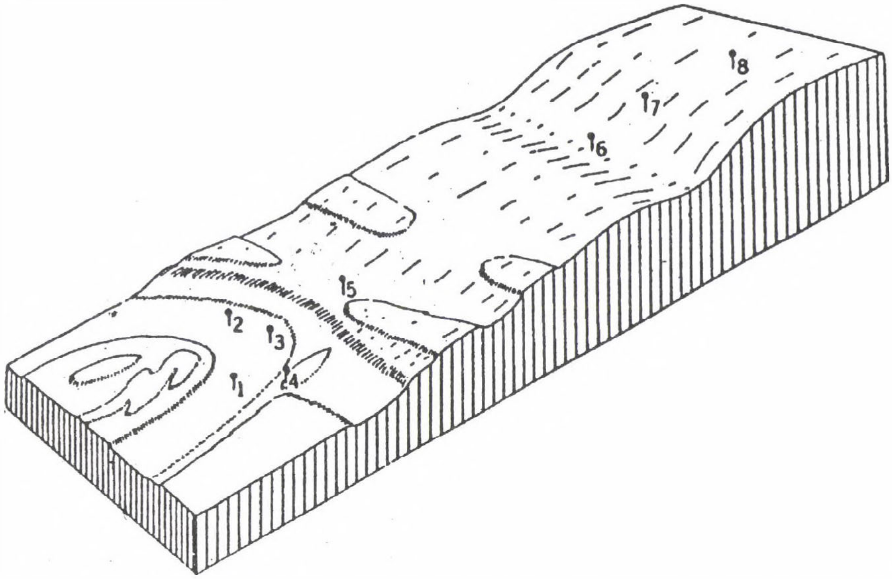

Fig. 2. Bloc diagram of the sites of microclimate measurements at Makád, from the natural biotopes of the ox-bow lake with seasonal waterlogging to the agrogeneous ecotopes, with the indication of the sites of measurements (1–8). For the explanations of 1–8 see the text

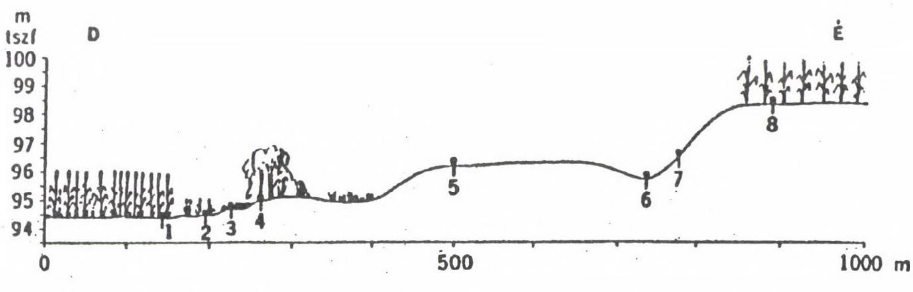

Fig. 3. Cross-section of the microclimate measurement sites from the natural biotopes of the ox-bow lake with seasonal waterlogging to the agrogeneous ecotopes, with the indication of the sites of measurements (1–8). For the explanations of 1–8 see the text

of a developed stand, the latter make the otherwise barren, treeless landscape more vivid. Their preservation is especially desirable with regard to the touristic potential of the lengthy section bordering the Danube. The hay meadow with high sedge and the reed margin are best used as meadows. However, it should be taken into consideration that sedge is unfavourable as fodder for cattle (it hurts the tongue of the animal) and gradual meadow improvement is necessary, first of all, with the introduction of leguminous crops.

Touristic functions along the Ráckeve-Danube are expected to be expanded further and this requires the infilling of lower-lying strips and agriculturally less valuable depressions.

Flood control measures and the resulting spread of cultivation induced a desiccation trend in soils and the genetic soil type has transformed into meadow soil with chernozem dynamics here, too. Further farming activities including frequent deep ploughing and loosening might help reduce compaction caused by the high colloidal content and a more favourable soil structure can be achieved. The terrace chernozem and lowland chernozem soils at stations 5 and 7, respectively, are more suitable for crop cultivation than the above ones. The extremely high $CaCO_3$ content in C horizon and traces of oxidation and reduction processes are counterbalanced by the appropriate depth of humous layer. Since $CaCO_3$ content is also high in the humous layer, due attention should be paid to this circumstance in the application of fertilisers.

In contrast, the cultural chernozem of station 8 is unsuitable for crop cultivation both for its low humus and high $CaCO_3$ contents as it is attested by the poor quality of the maize produced. For amelioration acidising fertilizers, for the recharge of nutrients green and stable manuring and conservation against erosion and shallow ploughing are needed.

Conclusions

Space as the subject of ecological investigations, is constituted of mosaic elements arranged horizontally and vertically, which can be integrated in accordance with the purposes of academic or practical evaluations, but the time factor (to the shortest time units) has always to be taken into account. Processes take place in temporal sequence, comprising natural phenomena and the social interference (of them the most important are regulation of the Danube and the resulting changes in the water budget of the area) leading to the present situation. It is logical to include annual or seasonal changes in vegetation cover of agricultural land, in stands and soil conditions etc. involving changes in the total agrogenic landscape. The consideration of all these minor alterations is beyond the objectives of a rational survey of academic or practical aim. However, it is to be emphasized that similar investigations are indispensable for large-scale hydrological projects.

The results obtained in test areas of various ecological endowments can be extrapolated to the cultivated land of the whole country. The summary and generalisation of previous studies combined with further research is a task related to agroecological microregionalisation.

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ISSUES OF LANDSCAPE CHANGE IN HUNGARY: AN EVALUATION OF IMPACTS OF LAND PRIVATISATION AND RECLAMATION

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Abstract

The functioning of landscapes is highly dependent on land use types, their spatial pattern and dynamics. The paper focuses on two recent man-induced changes of the cultural landscapes in Hungary: land privatisation and the concomitant spreading of small-scale farming (on the example of selected former cooperative farms) on the one hand and reclamation of mining and industrial areas (on the example of the Mecsek Mountains black-coal mining region) on the other. The above interventions do not only affect land use but involve alterations in a range of landscape elements and, through internal and external interactions, control landscape functioning. The transforming field pattern in rural environments and the restoration of derelict land in urban and industrial regions is not merely an economic problem. The process and its outcomes can be evaluated from a landscape ecological viewpoint, ie. to what extent the resulting pattern is functionally efficient in the context of the broader region as well as how it satisfies social requirements.

Introduction

In the study of cultural landscapes the 'natural' functions of the landscape cannot be separated from 'landscape potentials', ie. its resources for the satisfaction of the needs of human society (MAROSI, S. 1990; BASTIAN, O.–SCHREIBER, K.-F. 1999). One of the most comprehensive listings of landscape functions (BASTIAN, O.–RÖDER, M. 1997) includes a series of partial functions under three main headings: productive, regulatory and (hardly quantifiable) social functions *(Table 1)*. A starting point for the evaluation of any human intervention should be whether its eventual outcome is considered an improvement or a deterioration of the fulfilment of all or most of the listed functions.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T 016723, T 030859.

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Table 1. Major functions of a landscape (revised after BASTIAN, O. 1997)

Functions	Productive (economic) functions	Regulatory (ecological) functions	Social functions
Principal	ensuring renewable resources	regulating matter and energy flow	psychological
Partial	biomass crops (arable, plantations, meadow and pasture management) forestry animals (stock breeding, game management, fishery)	soil functions protection against hazards (erosion, waterlogging, desiccation, compaction) decomposition of materials introduced to the soil filtering, buffering and transforming functions	aesthetic function (landscape beauty) ethical function gene stock, historical landscape, cultural heritage
	Conservation of (surface and subsurface) water resources	hydrological functions (groundwater recharge, runoff regulation/retention, recovery of surface water quality	information, scientific functions research and education environmental damage
	ensuring non-renewable resources	meteorological functions (temperature moderation, increasing air moisture, controlling air motion)	human ecological functions bioclimatic impacts, filtering and buffering functions, chemical impacts (on soils, water and air)
	mineral resources extraction (ores, non-metallic minerals, building material quarrying)	regulating and regenerating populations (filling ecological niches) maintaining species and morphological biodiversity, habitat function	acoustic impacts (noise prevention)
			recreation function a combination of psychological and human ecological functions

Changes in rural land ownership and cultivation

Similarly to other countries in the eastern half of Europe, agriculture in Hungary underwent several *radical changes* in the 20th century. In the 1950's and early 60's peasants were deprived of their lands during the various waves of forced collectivisation (formation of cooperative farms). In 1991, however, when the transformation of the political regime in Hungary reached agriculture, a series of acts of compensation were passed by Parliament. The peasants of the post-war period were offered compensation vouchers by authorities and could bid for land at auctions organised for them, without being restituted to the property they lost. The members of cooperative farms received their share from the theoretically 'common property'. The cooperative farms had to identify tracts of land available for privatisation and allotment among the new owners. In spite of heavy criticism, the universally acknowledged measure of land quality was the modified gold-crown, first introduced in 1875.

A key issue in the reform of Hungarian agriculture has been *farm size* (HANTÓ, Zs. 1994). The land privatisation policy focused on the fragmentation of jointly cultivated land and neglected existing trends in agriculture, viz. the household plots, which could have developed into private commercial farms. A vast majority of the almost two million new landowners are unwilling or lack the necessary skills to cultivate their land on private farms (KESERŰ, J. 1993). Now the two million people own more than 8,000,000 plots of almost 6,000,000 ha total area. Also market conditions are unfavourable for agricultural produce and small-scale farming is usually on the margin of profitability. This explains that private (family) farming has been attempted at on less than 30 per cent of land (KOVÁCS, T. 1996). Although a spontaneous process of concentration both in the ownership and cultivation of land is observed, there is no indication of a short-term consolidation in agricultural land use pattern.

Focusing now on landscape ecological aspects, it is claimed that the human interventions are assessed positive or negative solely on the basis of their impact on landscape functioning. Although agroecosystems are strongly inhibited in their self-regulation, they fulfill vital functions (FINKE, L. 1986; HABER, W.–SALZWEDEL, J. 1992, BASTIAN, O. 1997): In addition to their major role of supplying plant and animal produces, ideally, they are important in nature conservation (preserving biodiversity), recreation (allowing part-time farming as a leisure activity), water management (maintaining soil moisture status and protecting surface water quality) and climate control (creating a pleasant microclimate). A serious criticism directed to large-scale farming (MÁRKUS, F. 1992) was that, overemphasizing the production function, it is unable to fulfill all these requirements in a harmoniously balanced way (Van MANSVELT, J. D.–MULDER, J.A. 1993). The EU agricultural policy serves ecological purposes (VERBAKEL, A. D. *et al.* 1984), along with the reduction of overproduction, through the introduction of extensification benefits (BALDOCK, D.–BEAUFOY, G. 1993). In Hungary too the interdependence of agriculture and nature conservation is increasingly underlined (HARRACH, T. 1992; IUCN 1992; TARDY, J. 1994) and a limit is formulated to intensive farming (ÁNGYÁN, J. 1991, 1993; ÁNGYÁN, J. *et al.* 1998): 8 to 12 per cent of agricultural land should be occupied by (semi)natural ecotopes in order to preserve something of the traditional pattern of rural land.

In principle, the transition from large-scale and intensive farming to small-scale private farming is regarded ecologically a promising trend. The following *advantages* can be expected as small-scale farming is spreading:

– the 'impoverished' landscape pattern of large fields regains some of their initial diversity;

– with more human labour employed, lesser amounts of chemicals are necessary;

– new, ecologically valuable ecotones are created along field boundaries;

– erosion hazard is decreased through the reduction of slope length;

– the aesthetic quality of the landscape is enhanced (new hedgerows, tree rows, terraces etc.).

Recently, a survey (LÓCZY, D. et al. 1999) was conducted to disclose whether the above advantages of small-scale private farming are already recognisable in the rural landscapes of some test areas in Hungary. The most difficult problem was to find a more or less exact procedure to decide whether a newly emerging pattern of land use is more favourable from landscape ecological aspects than a previous one.

The method applied involves a double approach. The transformation of land use pattern is evaluated on the basis of

1 the 'natural physical texture' of the landscape and

2 the historically evolved traditional pattern.

The first approach requires the compilation of knowledge on the physical components of the landscape (ÁDÁM, L. et al. 1981) and their interpretation for homogeneity or heterogeneity. Soils being the best indicators of ecological properties in the landscape, their spatial distribution can be regarded the truest reflection of 'physical texture'. The fuzzy nature of soil type boundaries (YEE, L. 1987; BURROUGH, P. A. 1989), however, make the comparison with land use (rather characterised by crisp sets to which Boolean operations apply) difficult. Topographic diversity provides rough guidelines for reconstructing landscape pattern.

The latter approach involves the critical analysis of historical land use surveys and interpretation of land use pattern on their basis. Since social factors have played a decisive part in the development of this pattern, this historical approach only provides very general guidelines for the evaluation.

The areas of four former cooperative farms, differing in physical endowments (in floodplain, foothill and blown-sand environments – MAROSI, S. – SOMOGYI, S. 1990) and in the level of farming, were selected for the study. The findings show that landscape pattern has not yet changed as large-scale farming of privately owned plots still prevails. The small plots designed improperly promote neither economic nor ecological functions. The observations in test areas suggested that it takes a rather long time until actual land use stabilises and a stable pattern develops.

The need for a rationalisation of field pattern was clearly visible just in the first phase of privatisation (KNEIB, W.–KURUCZ, M. 1996). In addition, in 1999 scandals broke out around the controversial re-purchase of previously privatised land valuable for nature conservation as well as the design of agricultural field boundaries was assessed improper in respect to the drainage of excess water (occupying hundreds of thousands of hectares in Hungary). They all highlighted the deficiencies of the land privatisation scheme, partly inherent and partly due to inappropriate implementation.

Land reclamation in collieries

If agricultural land use can be criticised for an insufficient consideration of the natural conditions of the land involved, abandoned mining landscapes are really the archetypes of destroyed land. Although land restoration affects much more restricted land areas, it also provides an opportunity for true restoration, ie. to design a cultural landscape closer to natural conditions at the same times fulfilling a series of human needs. With the decreasing demand for, reduced competitiveness and, thus, the inevitable closure of deep mines and abandonment of sections of open pits, an opportunity is provided to restore spoil tips and mining estates, in many cases close to urban areas, and to find land uses which serve the interests of local population (HESTER, R. E.–HARRISON, R. M 1994).

In the *Mecsek* Jurassic *black-coal mining* area (SZIRTES, B. 1994) five major open-cast pits and a series of waste tips stand before reclamation. In most of them waste was accumulated irrespective of the previous topography and drainage of the area (ERDŐSI, F. 1987). Therefore, technical reclamation measures are inevitable. There is a variety of impacts waste tips exert on their broader environment (LEHMANN A. 1980; CZIGÁNY, Sz.–LOVÁSZ, Gy.–VARGA I. 1997). In addition to the remediation of the damage caused by the poor design of tips, the landscape pattern has to be restored. This does not only involve the creation of optimal slope and drainage conditions but the careful adjustment of land use pattern to the broader environment.

One of the focal areas of reclamation planning is the Pécsbánya open cast (TOTAL Kft. 1997) of 47 ha area (11 ha internal tip and 36 ha external tip). The volume of the void is ca 15 million m^3 and the pit is 100–150 m deep relative to the internal tip. The mining area is limited by the forest belt of residential areas and waste tips and, thus, further expansion is not possible. Also the Pécs Power Plant Co. was unsuccessful to obtain permission for opening a further open cast, which could have supplied waste in sufficient amounts to fill in the existing void. The first versions of the restoration plan, which covered the mining estate and its broader environs (ca 300 ha area) have to be revised completely. Given the proximity of the mine to the centre of the city of Pécs, however, a very careful planning of reclamation measures is needed. The demands for green belts, residential and recreation areas and communication routes have to be considered. With the rising prices of building plots on the southern slopes of the Mecsek Mountains within the city boundaries, there is a financial pressure to allot as large an area for housing developments as possible. However, the interests of the whole city require more green belts (amenity forests) to improve air quality, one of the poorest in Hungary.

The surveys of the environmental geological conditions in the area show that they have changed fundamentally as a consequence of mining. The disturbances caused in stratification rearranged groundwater flow regimes (TOTAL Kft. 1997) and, thus, surface stability may not satisfy the requirements of housing development. If – as a consequence of the final closure of open-cast mines – filling materials will not be available in sufficient amounts, groundwater may rise to disadvantageously high levels and may cause ponding in the present void area. The high sulphur content of the Mecsek black coal (A mecseki ... 1994) also poses a problem. Anyhow, it seems to be certain that housing development cannot take place immediately after the landscape restoration measures. Observations have to be made for several years to estimate the rate of compaction, an adjustment to the rapid intervention of waste accumulation. It is a common characteristic with land reprivatisation that the impacts of radical changes in landscape pattern only manifest themselves at a later date.

The restoration of the topography prior to mining activities could be achieved with some restrictions. The initial surface was dissected by stream valleys into three or four low ridges of north–north-west to south–south-east strike. This pattern also controlled the location of forests and meadows. Since the creation of extensive surfaces, available for a wider range of uses, enjoys priority, the restoration of all of the valleys to their original shapes in the waste tip area does not appear among the goals of the reclamation plan. The tip will retain its terraced conical shape of the time when mining was abandoned. Although infiltration is expected to be of high rate in the unconsolidated fill, for erosion control it is essential to reduce the lengths of uniform slope segments below ca 100 m. In contrast, downslope the waste tip two valley heads of 4–5 per cent side slopes and ca 2.5 per cent slope along the thalweg are planned to be formed. To promote revegetation (afforestation or grassing), the application of mycorrhizal inoculation techniques based on international experience (STURGES, S. 1997) is advised. The valley heads will be environments of favourable microclimate in sheltered position. Therefore, in addition to utilisation as amenity forests or grasslands, recreational uses are also envisaged for them.

This example of land restoration clearly illustrates the range of requirements set against land reclamation planning. A short summarisation of those requirements could be the following. The natural pattern of the landscape should be retained as precisely as possible and the harmony between landscape elements should be ensured. Such considerations may increase the efficiency of landscape functioning.

Acknowledgement – Author is grateful for a four-year research grant from the Hungarian National Science Foundation (OTKA, ID number: T 16 723) and for Pécs Power Plant Co. Ltd. for giving permission for the use of manuscript coal mine reclamation plans.

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Studies in Geography in Hungary 32, Geographical Research Inst. HAS, Budapest 2000, pp. 169–178.

THE MEDVES PLATEAU

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In the area of the North Hungarian Mountains, between the Danube and the Bodrog River there can be found andesite volcanic masses (e.g. Börzsöny and Mátra Mountains), and mountain blocks composed of karstic limestones (e.g. Bükk Mountain) as well as dissected low hills and basins built up of sedimentary rocks. Among them the young Nógrád–Gömör Basaltic Area covering about 430 km^2 is situated along the Hungarian-Slovakian boundary,. Two thirds of it belong to Slovakia and one third, the southern part, to Hungary. One of its most conspicuous parts, just dissected by the state boundary, is the Medves Plateau (in short "Medves"), the only basaltic lava plateau of Hungary *(Fig. 1)*.

The geological setup of this basaltic area is varied. Superimposed on the little studied Proterozoic-Paleozoic crystalline basement, mostly near-shore neritic sediments are dominant as a consequence of transgressions having begun about 40 million years ago. Within the following 20 million years, from the end of the Eocene till the beginning of the Miocene, extension and depth of this sea had been in a permanent change entailing deposition of bottom sediments in various, locally vast thicknesses (mainly clay marls, marls, sandstones and very fine-grained aleurolites /schlieren/). About 20 million years ago, simultaneously with the strong uplift of the Carpathian Mountains, the sea regressed and a varied topography emerged with intermontane basins. Some basins were occupied by swamps with lush vegetation, from which brown-coal beds originated. In the Ottnangian intense volcanic activity began with heavy explosions, spreading out thick rhyolitic and rhyodacitic clastic rocks (mainly ignimbrites). During the Karpatian and mainly in the Badenian andesitic lava flowed onto the surface. In some places the magma had not reached the surface, so laccolithes were formed and later outcropped; among them the highest is the Karancs (729 m) lying on the western rim of the basaltic area.

Significant changes occurred again during the Pliocene. Owing to the steadily strengthening tectonic movements the area had become dissected, certain tectonic units tilted asymmetrically along the faults; some of them lifted up while others subsided to

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Fig. 1. Map of Medves and its surroundings

different extent. Beginning 6.5 million years ago and lasting till the end of the Late Pleistocene, heavy volcanic activity took place again in several phases, which can be explained by the anomalous thin crust, the high heat fluxes and the extensional stresses. In consequence of these factors melted material of the upper mantle came up along the deep faults of the crust and appeared as basalt volcanism. Both lava and pyroclasts came to the surface and about 100 separate volcanic formations came into existence, among them cones, lava plateaus, explosion craters (tuff rings, maars) and smaller outweathered dykes and necks.

As regards the eruptions, little regularity can be observed. Examples of eruption in one or more phases, or of formation of both lavas or pyroclasts can be found. The lavas are mainly compact, in some places blistered or scorious. The prismatic structure of beautiful, mainly pentagonal or hexagonal columns, 10–25 cm in diameter, frequently occurs. The lava plateaus are bedded and laminated. In the stratovolcanic activity no general regularity concerning the alternation of explosive and effusive activities appears, although in the course of the initial activity the former is more frequent. Also postvolcanic activity can be traced, mainly in the northern part of the area. There the lava plateaus (Bucsony, Monosza, Pogányvár, Medves) are more typical, while in the south smaller cones (Salgó, Nagy-kő, Somos-kő) are met.

Following this volcanic activity, the topography had undergone strong alterations during the Quaternary having tectonically and erosionally dismembered owing to the young and quick uplift (200–350 m) and of the climatic changes. In the Pleistocene, especially during glacial phases, disintegration of rocks, solifluction, soil creeping and other mainly derasional mass movements took place, while in the pluvial periods weathering and ablation played significant role. At the foot of the basaltic landforms block fields can be met frequently, consisting of boulders and pieces of basaltic columns, created mainly by periglacial attrition. The lava plateaus, e.g. the Medves receded parallel with their rims. In some places slides and other mass movements of large extension took place, playing important role in the geomorphic evolution.

During the Holocene man-induced effects caused considerable changes in relief modelling *(Fig. 2)*.

As it was mentioned above, the Nógrád–Gömör Basaltic Area consists of some small lava plateaus (among other landforms). One of them is the Medves, extending 6–7.5 km north-south, 2–3 km east-west, covering 12.8 km^2, from which 7.8 km^2 belongs to Hungary. Its lowest portions are about 500–525 m a.s.l., the average altitude is 550–570 m. It is really a plateau, especially its southern part is nearly table-smooth where the inclination is only some per mille. In the middle of the plateau the 'Medves magosa' rises abruptly above its surroundings, with the uppermost part of the whole plateau (658.6 m) on its flat crest *(Photo 1)*. North of it the surface is more rugged than in the southern Medves. The relative relief is on the average only 45–50 m/km^2, 70 m/km^2 on the south, but more than 100 m/km^2 on the north. The petrographical boundary of the basaltic rock is a very sharp morphological marker as well, therefore the plateau rises with steep rims above its surroundings area, especially on the east, with 300 m difference of level, where the angles of slope are 20–40° on the average.

The basement of the plateau is constituted by Oligocene-Miocene sediments, e.g. the Szécsény Schlier Formation, which consists of aleurolite, the strongly cross-bedded Pétervására Sandstone Formation, with a lot of glauconite and many loaf-like concretions bulging out, the ignimbrite-like Gyulakeszi (so called "lower") Rhyolitic Tuff Formation, with truly rhyodacite ignimbrites, created by heavy explosions, and the Salgótarján Browncoal Formation, which had been exploited and is originated from the lush vegetation of swampy bays. The beds of these formations are

Fig. 2. Geomorphological map of the Medves Plateau. (Ed. by PINTÉR, Z.) – 1 = stable slopes; 2 = slopes with landslide hazard; 3 = summit level in higher position (300 m); 4 = summit level in lower position; 5 = higher crest (300 m); 6 = mountain ridge, interfluvial ridge; 7 = gentle slope segment; 8 = hilltop; 9 = saddle; 10 = flood plain floor; 11 = erosional ravine; 12 = erosional gully; 13 = streambed with steep slopes; 14 = deep (20 m) erosional valley; 15 = flat and wide erosional valley; 1 6 = erosional-derasional valley; 17 = derasional valley; 18 = derasional niche; 19 = settlement

Photo. 1. Medves Plateau, with the uppermost part of Medves in the background

relatively horizontal, they have only a slight dip southward and are cut tectonically by parallel and perpendicular faults, forming a "horst-graben-type" geological structure.

The young Salgóvár Basalt Formation lies with a considerable hiatus unconformably on these sedimentary rocks. The basalt volcanic activity here took place about 2–2,5 Ma BP. In spite of the fact, that both the petrography and the stratigraphy of the plateau is studied quite well, neither the exact process of the eruptions nor the place of the eruption centre or centres have been cleared up yet. Data given by the coal-miners prove that at the beginning of the basalt volcanic activity the topography could be very rugged not only owing to the tectonic dissection but also because of strong erosional processes taken place on the basement of the area earlier. This rugged topography had been levelled off by the overburden basalt and the basaltic "cap" protecting the underlying sedimentary rocks have recently become outcropped from its vicinity.

The valley network on the plateau is rather scanty. On the southern part broad dells prevail, whereas less erosional valleys can be found. These erosional valleys are generally straight, narrow and shallow, and can be found in the line of intersection of slopes having different dip, which suggests tectonic origin. There are also exceptions, such as depressions induced by human activities, or some regressional valleys originated from the rims. A more extensive valley network exists on the north, but altogether the valley density (excluding the deep valleys of the rims) remains less than 1 km/km^2. On the contrary, the steep rims are strongly dissected *(Photo 2)*. Here slow withdrawal can be verified, but not so intense as earlier was assumed by researchers due to the intense uplift having taken place since the Pleistocene. There are narrow and very deep

Photo. 2. Erosional gully of man-made origin at the eastern margin of Medves

Photo3. Waterfall of Gortva Stream

regressional valleys developed from steep gullies especially on the north and north-east. Particularly beautiful is the valley of the Gortva River with its spectacular (1.7 m broad and 2.7 m high) waterfall, where the water rushes down on a thick, hard sandstone bed *(Photo 3)*. This Gortva Valley should withdraw "only" 500–700 m yet for cutting the Medves and so to separate a smaller (1–1.5 km^2) part of the plateau. Most of these valleys on the rim are asymmetrical showing the structure of the underlying (and southward dipping) strata but the stronger frost shattering on the slopes of northern exposure is partly responsible for it.

In the Pleistocene the topography has been strongly modelled by periglacial processes, above all by mass movements. Also nowadays the attrition and particularly the mass movements on the slopes (creeping, slides) are very intense processes on the plateau. These movements can be studied above all also in the eastern rim, especially on the sides of the above mentioned deep erosional regressional valleys, where young slides can also be seen. The character of some geological strata (mainly the variegated clay and rhyolitic tuff) and the slight dip of the sedimental rocks was often favourable for bursting out them.

Apart from the valleys and from the height of the "Medves magosa" the topography of the plateau is rather uniform. On the mainly table-smooth, partly slightly rolling surface only small closed drainage basins can be found. In these wet, swampy depressions the vegetation have specific features. In various places birch groves form interesting circular shape; probably indicating former depressions which have already filled up. Some depressions have tectonic origin, but most of them are man-made.

As regards the morphology of the Medves it is unique that human activities have played an eminent role in geomorphic evolution. Especially the landscape forming effects of the very intensive quarrying of basalt and the extraction of the underlying brown-coal strata, and that of the related industries and infrastructure shaped the surface and the natural vegetation. Coal mining in this region was very important for Hungary, especially after the Trianon Peace Treaty, when Hungary lost almost all of its mines in the Carpathian Mountains. The considerable development of infrastructure (inclined shafts, bogie-tracks, transmission lines, narrow-gauge railway tracks, ropeways, cuts, ramparts, tunnels etc.), and also the big cicatrices caused by the mining, the enormous refuse dumps, have significantly altered the surface as well. These processes have been continued after the Second World War, when the communist rule wanted to convert Hungary into a "the country of the iron and steel", without respect of the real physical potential (and of the very law calorific value of these brown-coal beds). Because of the geological set-up the beds (with the brown-coal strata lying under the basáltic surface), the mining galleries extended to a great distance inside the plateau.

The first basalt quarries were opened at the end of the 19th century because of the great demand on building materials and due to the good quality of the basaltic rocks on the Medves Area their number increased quickly. There was an especially intense working on the rims of the plateau. It was interesting, that the mining of the two kinds of raw materials (basalt and coal) had used the mining infrastructure jointly for a long time. In the 1970's, however, because of the bad economic efficiency indices almost all of these mining activities – both the coal and basalt mining – became abandoned. After cancelling subsurface mining under the plateau and drawing off the wooden supports the galleries collapsed and oval, rounded, and irregular-shaped or longitudinal, trench-like depressions appeared on the surface. The most spectacular man-made landform produced by undermining can be seen very close to the Medves. On the top of an other eruption centre (Szilvás-kő) several tens of metres long, earlier in some places 20 m deep, nowadays 5–10 m deep trench-like fissures can be observed. These fissures have partly been filled up and sometimes continue in conspicuous deep caves.

The curved, so called "drunken" trees and the recent fissures suggest that these movements are still active. Along such a fresh fissure an interesting phenomenon took place in 1995–1996, when in the depth, within the brown-coal strata spontaneous ignition occurred: due to the fresh air inflow the burning was strong, the fissure smoked, by their sides the snow had melted and because of the special microclimate a unique vegetation has developed.

Also the future of the abandoned basalt-mines poses considerable environmental problems. Once operating, now abandoned open-cast mines extend over 52 ha and there are refuse dumps over 30 ha on the Medves. These damaged surfaces have not been reclaimed and nowadays they are often occupied by illegal waste deposit sites. Another environmental problem is, that former workings, especially the waste dumps are covered either by scant vegetation, or on the contrary by dense, hardly penetrable scrub.

The climate of the plateau is moderately cool and moderately dry. Annual totals of sunshine hours are rather low (about 1850–1900 hours) just like the annual mean temperature (about 7–8 °C). The annual mean precipitation is about 600 mm, but its temporal and spatial distribution is uneven. Northwestern and western wind directions are prevalent.

According to the floristic-phytogeographical divisions the plateau belongs to the Pannonicum flora province, within that to the Matricum flora sector and to the Agriense flora district. Originally the plateau was densely covered by oak and beech forests. In the 19^{th} century because of the high timber consumption (e.g. railroad construction) these forests were cleared. Also the mining activity needed a lot of wood. An other modification of the natural vegetation was caused by mistaken agricultural activity (on the plateau once there were hayfields, pastures, and extended arable lands, too). Later these activities were given up (but animal grazing sometimes is observable) partly owing to the low fertility of soils (most frequently they are brown forest soils with clay illuviation and brown earths). Nowadays on the proper place of the former forest clearance shrub (e.g. blackthorn, hawthorn, and particularly blackberry) grows. There are also scattered birch groves and along some valleys remnants of the original oak forests have survived. In spite of the changes the Medves has fortunately a unique vegetation, with some rare species (e.g. Dentaria bulbifera). Some species of protected animals live in the area, as well, e.g. bull-frog, salamander, kingfisher, big dormouse, buzzard.

Because of the geological and geomorphic uniqueness and of these natural values the whole area of the Medves is protected as part of the Karancs–Medves Landscape Protection Area. The first part of this area was established in 1964, on 129 ha, comprising the basaltic cones of the nearby Nagy- and Kis-Salgó. By 1989 the protected area had already extended to 7000 ha, with 450 ha falling into the strictly protected category. The Nature Conservation Office which is responsible for the management of this protected area had stairs and banisters built up, put out educational tables for the public. Unfortunately, the insane vandalism often runs counter these efforts. Nevertheless there are plans for creating new study-trails, e.g. a cycle track will be created along the old track of the abandoned narrow-gauge railway line which was used for transportation of the coal and basalt and along this cycle track educational tables will be put out. From the aspects of nature conservation it is very important that at the northern continuation of the Karancs–Medves Landscape Protection Area, on the Slovak side there is an other conservation area (Chránená Krajina Oblast' Cerova Vrchovina) with similar to the Medves lava plateaus (e.g. Pogányvár, Monosza) and within that there are also strictly protected parts (e.g. the spectacular Somos-kő). It is to be expected that a joint protected area will be established soon hopefully providing excellent opportunities also for geographic research.

This publication was supported by the National Scientific Research Fund (OTKA), Project No. T017824.

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